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THE QUATERNARY ERA
WITH SPECIAL REFERENCE
TO
ITS GLACIATION

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By

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To the Memory of

PROFESSOR P. F. KENDALL, F.R.S.

Illustrious geologist, inspiring teacher, generous friend

"Ages are spent in collecting materials, ages more in separating and combining them. Even when a system has been formed, there is still something to add, to alter, or to reject. Every generation enjoys the use of the vast hoard bequeathed to it by antiquity, and transmits that hoard, augmented by fresh acquisitions, to the future ages. In these pursuits, therefore, the first speculators lie under great disadvantages, and when they fail, are entitled to praise."

LORD MACAULAY. *Essay on Milton*

"*Wer kann was Dummes, wer was Kluges denken*
Das nicht die Vorwelt schon gedacht?"

GOETHE'S *Faust* (Part II, Act II, Scene I)

PREFACE

"When the work of the geologist is finished and his final comprehensive report is written, the longest and most important chapter will be upon the latest and shortest of the geological periods" (G. K. Gilbert, *612*, 1). This somewhat hackneyed phrase still expresses an important truth; for this final chapter tells of the most fascinating and dramatic of all geological events. The Quaternary saw greater changes than probably any other span of equal length in the earth's history; interwoven with it are the later development of the biosphere and the origin and progress of the human race.

Quaternary geology is a study of little more than one hundred years yet its literature is mountainous. Alone of the geological formations it has its own journals, the *Glacialists' Magazine* (1893–7), *Revue de Glaciologie* (1901–7), *Zeitschrift für Gletscherkunde* (1906–44), *Zeitschrift für Gletscherkunde und Glazialgeologie* (1947→), *Bollettino del Comitato Glaciologico Italiano* (1914–41), *Quartärperiode*, Kiev (1930→), *Zeitschrift für Geschiebeforschung* (1925–42), *Die Eiszeit* (1924–30) *und Urgeschichte* (1930→), *Quartär* (1938–41), *Eiszeitalter und Gegenwart* (1951→)—this continues *Quartär* and *Zeitschrift für Geschiebeforschung—Jökull* (1952→), *Journal of Glaciology* (1947→) and *Quaternaria* (1954→) It also has had its own international congresses (Copenhagen, 1928; Moscow, 1932; Vienna, 1935; Rome and Pisa, 1953). The extension into cognate fields has led to the wide dispersal of papers and memoirs from these and purely geological journals into innumerable journals, magazines, jubilee publications and books, into the proceedings and transactions of academies and scientific societies, geographical, geomorphological, palaeontological, zoological, botanical, ornithological, pedological, astronomical, archaeological, anthropological, anatomical, meteorological and oceanographical, and into publications of universities, government surveys and official or private expeditions, many of them difficult of access. Additions, made at an ever-increasing rate and over an ever-widening field, appal the keenest student who is unable to keep abreast of the rising tide of detailed and specialised information. Valuable observations and deductions are entombed as soon as entomed.

Searching in this gigantic literature is most laborious and involves much barren and repetitive reading. The language difficulty is perennial and present methods waste time, effort, money and temper. There is therefore an urgent need of an objective work collecting and correlating this heterogeneous material and embodying in reasonable compass a digest of the imperfectly organised data that have accrued from the labours of generations of workers in the Quaternary field. When brought into juxtaposition with related facts in the mind of the observer, these stores

of useful information may stimulate the growth of knowledge and the birth of new ideas.

Perhaps no geological period has so divergent views as has the Pleistocene. Indeed, Quarternary geologists have long enjoyed the unenviable reputation of being among the most disputatious. The Quaternary is the battle-ground of rival views held by geologists whose opinions are entitled to respect. In numerous cases of conflict of fact or opinion it is beyond the author's powers to resolve them; for the facts given in a book of this kind can be based only to a slight extent on personal acquaintance. I have endeavoured to give an objective presentation of the present position on the various Quaternary questions and an orderly analysis of pertinent knowledge welded into a coherent whole. What the book loses in readability by its fullness of treatment and severe economy of words it will I hope gain in usefulness.

The progress of science involves a succession of hypotheses; many die in infancy and few survive. Some would doubtless never have been born had the pioneers had more facts. Many theories therefore have now only historic interest, though few are without a germ of truth. Even exploded theories have a habit of being resurrected. This is perhaps inevitable where the processes are so often hidden from view and where so much is left to deduction and is admitted or rejected according to personal predilection or experience. Argument often leads only to probable conclusions. On many fundamental questions we cannot yet pronounce final judgment; the problem of ice-flow, the extent and manner of ice-erosion, the mode of deposition of till and drumlins, the altitude of land and sea during Quaternary time, the cause and mechanism of the epiglacial uplift, the number and lengths of the glacial and interglacial and pluvial and interpluvial epochs, and the evolution and development of man throughout the world are but a few examples. Indeed, much of Quaternary geology is still in the stage of multiple hypotheses, and the fate that has overtaken so many geological fictions awaits some which are widely current or held to be impregnable at the present day. The advanced worker, however, cannot dispense with such doubts, uncertainties and contradictions.

It may be said with some show of truth that I have laid too much stress upon the historical aspects and that discarded theories might well have been omitted. In view, however, of the fact that old ghosts are ever ready to reappear—the glacial submergence of Britain, the tectonic origin of fjords and formation of drumlins by wind and rain are three recent instances—one cannot be sure that any hypothesis has been finally relegated to the geological dustbin. Moreover, it is advisable that the present-day student, who has but an imperfect realisation of the obligations due to those who laid the foundations, should be familiar with the historical weapons and arguments.

J. Geikie, in preparing the first edition of his *Great Ice Age* (1874), had to abandon his plan of preparing a bibliography since it would have required a volume in itself. To-day, after the lapse

of a further 80 years and with the vastly accelerated output of Quaternary literature, such a task is well-nigh impossible. I have therefore confined myself to compiling a list of over 1900 entries of the publications, not necessarily the most important or important at all, that I have found recurring in the writing of the text. These have been supplemented by numerous references at the end of the chapters. For papers in Slav languages I have unfortunately had to depend on reviews, summaries or abstracts in other languages. Apart from these, about 800 publications have not been available to me: these are referred to on the authority of others. I should be grateful to any correspondent who finds errors or misstatements or mistakes in references, many of which have doubtless slipped in despite my utmost vigilance.

The text was completed at the close of 1953. Later material has only been incorporated where it was found possible to do so without disturbing the reference sequences. The authors are given in brackets in the text and the references on pp. 1601, 1602.

"I have long discovered", wrote Darwin, "that geologists never read each other's works and the only object in writing a book is a proof of earnestness and that you do not form your opinions without undergoing labour of some kind" (*349*, I, 334). This book is the outcome of a systematic study pursued over many years and would not have been possible but for the help I have received from many quarters. It is a great pleasure to acknowledge the debt I owe to those many workers in different fields who have provided answers to direct questions about which I was in doubt. I am also most grateful to many librarians for unfailing help and kindness, in particular to Miss A. Barber of the Geological Society of London, Mr. W. D. Woodrow of the Royal Geographical Society, Mr. W. S. Warton of the Linnean Society, Mr. H. B. Rowbotham of the Geological Library and Miss Phyllis Edwards of the Botanical Library, British Museum (Natural History), Mr. G. Woledge of the London School of Economics, and to Mr. J. Graneek and Miss A. Megaw of Queen's University, Belfast. I also wish to express my thanks to Professor J. Heslop Harrison, Queen's University, Belfast, who has given me invaluable assistance on the biological side, to Dr. W. Schwarzacher who read through the glaciological chapters and made many suggestions which I was glad to adopt, to Dr. R. Savage who has checked the zoological nomenclature, to Mr. A. Green, Ballycastle, Co. Antrim, who has checked the proofs, and to my wife and son Henry who have helped me with the references in typescript, proof and in other ways. I alone, of course, am responsible for any remaining obscurities and inaccuracies or errors of judgment.

With few exceptions the illustrations have of necessity been borrowed from published sources. The authors to whom acknowledgement is due are noted in the lists of illustrations in the text. For the loan of blocks or permission to reproduce the illustrations I wish to make grateful acknowledgement to the following for the text-figures given in each case: Akademische Verlagsgesellschaft, Lpz., 257, 271, 276, 313; American Anthropological Association,

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Ballycastle,
Co. Antrim,
Northern Ireland.
July, 1956

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Volume I

KEY TO ABBREVIATIONS OF TITLES OF PERIODICALS

Publications of societies and institutions have generally long titles. With few exceptions these have been greatly reduced in the references at the ends of chapters and in the Bibliography, Vol. II, pp. 1556–1601.

The general abbreviations used (unless modified in the list of publications set out below) are the following:

A.	*Association*
Ab.	*Aarbog*, etc.
Abh.	*Abhandlung*
Am.	*American*
Ann.	*Annals*, etc.
Ark.	*Arkiv*
B.	*Bulletin*, etc.
BB.	*Beilage Band*
Bbl.	*Beiblatt*
Bd.	*Band*
Bh.	*Beiheft*
Bot.	*Botanical*, etc.
Btr.	*Beiträge*
C.	*Congress*
CR.	*Comptes Rendus*
D.	*Deutsch.*
Erg.	*Ergänzungs.*
Erl.	*Erläuterung*
FC.	*Field Club*
Fbd.	*Festband*
Fschr.	*Festschrift*
G.	*Geology*, etc; *Gesellschaft*
Gg.	*Geography*, etc.
Gl.	*Glacial*, etc.
Gphys.	*Geophysik*
H.	*Heft*
Hb.	*Handbuch*
I.	*Institute*, etc.
int.	*international*, etc.
J.	*Journal*
Jb.	*Jahrbuch*
JB.	*Jahresbericht*
K.	*Karte*
Lb.	*Lehrbuch*
Lk.	*Landeskunde*
M.	*Mitteilung; Medd.*
Mo.	*Monthly*
nf.	*Naturforschung*
nh.	*Naturhistorisch*
nk.	*Naturkunde*
Nschr.	*Nachschrift*
nw.	*Naturwissenshaft*, etc.
P.	*Proceedings; Paper; Publication*
Pal.	*Palaeontological*, etc.
Ph.	*Philosophical*, etc.
PP.	*Professional Paper*
R.	*Review*, etc.
RC.	*Rendiconti*
Rp.	*Report*
S.	*Society*, etc.
SB.	*Sitzungsbericht*
Sbd.	*Sonderband*
Sc.	*Science*, etc.
Schr.	*Schriften*
Sv.	*Survey*
T.	*Transactions*
U.	*University*, etc.
Un.	*Union*
V.	*Verein*
Ver.	*Veröffentlichung*
Vh.	*Verhandlung*
vl.	*vaterländisch*
Z.	*Zeitschrift*
Zbl.	*Zentralblatt*

The titles of journals, etc., frequently used, have been much reduced. The more important abbreviations are set out below, the contractions on the right of the page being those sanctioned in the *World List of Scientific Periodicals* compiled by the World List Association (Ed. 3, 1952). For further abbreviations to the references, see pp. xxvii, xxviii.

A. J. S.	*Amer. J. Sci.*
Abh. bayer. Ak.	*Abh. bayer. Akad. Wiss. Maths. Phys. Kl.*
Abh. Gg. G. Wien	*Abh. geogr. Ges. Wien*
Abh. LA.	*Abh. preuss. geol. Landesanst. New Folge.* *Berlin*
Abh. preuss. Ak.	*Abh. preuss. Akad. Wiss.*

Act. S. Helv.	*Verh. Schweiz. naturf. Ges.*
Am. G.	*Amer. Geol.*
Ann. G.	*Ann. Géogr.*
Ann. Hydr.	*Ann. Hydrogr. Berl.*
Ann. Mag.	*Ann. Mag. nat. Hist.*
Ann. N.Y. Ac.	*Ann. N.Y. Acad. Sci.*
Ann. Obs.	*Ann. Obs. Mont Blanc.*
Ann. Phys.	*Ann. Phys. Lpz.*
Ann. S. G. N.	*Ann. Soc. géol. Nord*
Ant.	*Antiquity*
Anthr.	*Anthropologie. Paris*
Appal.	*Appalachia*
Arch.	*Archaeologia*
Arch. I. P. H.	*Arch. Inst. Paléont. hum.*
Arch. M. N.	*Arch. Math. Naturv.*
Arch. Meckl.	*Arch. Ver. Naturg. Mecklenb.*
Arch. Sc.	*Arch. Sci. phys. nat.*
Arch. Teyl.	*Arch. Mus. Teyler.*
Ark. Kemi.	*Ark. Kemi Min. Geol.*
B. A. Gg. S.	*Bull. Amer. geogr. Soc.*
B. Ac. imp.	*Bull. Acad. Sci. St.-Petersb.*
B. Ac. Pol.	*Bull. int. Acad. Cracovie (Cl. math. nat.)*
B. C. G. Petersb.	*Bull. Com. geol. St.-Petersb.*
B. S. belg. G.	*Bull. Soc. belg. Géol. Pal. Hydr.*
B. S. G. F.	*Bull. Soc. géol. Fr.*
B. S. nat. Mosc.	*Bull. Soc. Nat. Moscou*
B. S. neuch.	*Bull. Soc. neuchâtel Sci. nat.*
B. S. Vaud.	*Bull. Soc. vaud. Sci. nat.*
Ber. nf. G. Freib.	*Ber. naturf. Ges. Freiburg. i. B.*
Ber. sächs. Ak.	*Ber. sachs. Ges. Akad. Wiss. math. phys. Kl.*
Bih. Sv. Ak. H.	*Bih. Svensk Vetensk Akad. Handl.*
Boston S. Occ. P.	*Occ. Pap. Boston Soc. nat. Hist.*
Boston S. P.	*Proc. Boston Soc. nat. Hist.*
Bot. Not.	*Bot. Notiser.*
Btr. G. K. Schw.	*Beitr. geol. Karte Schweiz.*
C. Min.	*Zbl. Min. Geol. Paläont.*
Can. Mus. B.	*Mus. Bull. geol. Surv. Can.*
Can. Mus. Mem.	*Mus. Mem. geol. Surv. Can.*
Can. Rp.	*Ann. Rept. new ser. geol. Surv. Can.*
CR.	*C. R. Acad. Sci. Paris*
CR. C. Arch.	*Congr. int. Anthrop. Archeol. prehist.*
CR. C. G.	*int. géol. Congr.*
CR. C. Gg..	*int. géogr. Congr.*
D. Boden.	*Dtsch. Boden*
D. G. U.	*Danm. geol. Unders.*
Dschr. Ak. Wien	*Denkschr. Akad. Wiss. math. naturw. Kl.*
E.	*Eiszeit (u. Urgeschichte)*
E. & G.	*Eiszeitalter u. Gegenwart*
Ecl.	*Ecl. geol. Helv.*
Essex N.	*Essex Nat.*
Étud. gl.	*Étud. glaciol.*
F.	*Fennia*
F. F.	*Forsch. Dtsch. Wiss. Fortschr.*
F. G. P.	*Fortschr. Geol.*
F. Lk.	*Forsch. Dtsch. Landesk.*
Finl. B.	*Bull. Comm. géol. Finl.*

G.	*Geologist.*
G. Abh.	*Geogr. Abh.*
G. Ann.	*Geogr. Ann. Stockh.*
G. Anz.	*Geogr. Anz.*
G. Charb.	*Geol. Charakterbilder*
G. F. F.	*Geol. Foren. Stockh. Förh.*
G. J.	*Geogr. J.*
G. JB.	*Geogr. Jber. Öst.*
G. M.	*Geol. Mag.*
G. Mijnb.	*Geol. en Mijnb.*
G. Pal. Abh.	*Geol. Paläont. Abh. Jena.*
G. R.	*Geogr. Rev.*
G. Rd.	*Geol. Rdsch.*
G. S. A. B.	*Bull. geol. Soc. Amer.*
G. S. A. Mem.	*Mem. geol. Soc. Amer.*
G. S. A. Spec. P.	*Spec. Pap. geol. Soc. Amer.*
Gfys. P.	*Norsk. Vidensk. Ak. Geofys. Publikasjoner*
Gg.	*Géographie*
Gg. T.	*Geogr. Tidsskr.*
Gl. M.	*Glacialists' Mag.*
Him. J.	*Himalayan J.*
I. N.	*Irish Nat.*
I.N.J.	*Irish Nat. J.*
Ind. Mem.	*Mem. geol. Surv. India*
Ind. Rec.	*Rec. geol. Surv. India*
INQUA	*Int. Assoc. Study Quaternary*
J. Conseil.	*J. Conseil int. Explor. de la Mer*
J. G.	*J. Geol.*
J. G. S. Dubl.	*J. Geol. Soc. Dubl.*
J. Gl.	*J. Glaciol.*
J. LA.	*Jb. preuss. geol. Landesanst.*
J. RA(BA)	*Jb. geol. Reichsanst. (Bundesanst.), Wien*
J. R. A. I.	*J. R. anthrop. Inst.*
J. R. G. S. Ire.	*J. Roy. Geol. Soc. Ireland. Dubl.*
J. S. AC.	*Jb. schweiz. Alpenkl.*
Leop.	*Leopoldina. Halle*
M. Bad. LA.	*Mitt. bad. geol. Landesanst.*
M. D. G. F.	*Medd. dansk. geol. Fören.*
M. D. O. AV.	*Mitt. dtsch. ost. Alpenver.*
M. Gr.	*Medd. Grønland*
Manchr. Mem.	*Mem. Manchr. lit. phil. Soc.*
Matér.	*Matér. Étude de l'Homme*
Matér. Étud. Cal.	*Matér. Étud. Calam.*
Mém. Ac. imp.	*Mém. Acad. Sci. St.-Petersb.*
Mich. Ac. P.	*Mich. Acad. Pap.*
Mo. W. R.	*Mon. Weath. Rev. Wash.*
Mus. C. Z. B.	*Bull. Mus. comp. Zool. Harv.*
Mus. C. Z. Mem.	*Mem. Mus. comp. Zool. Harv.*
N.	*Nature, Lond.*
N. Dschr.	*N. Denkschr. schweiz. naturf. Ges.*
N. G. T.	*Norsk. geol. Tidsskr.*
N. G. U.	*Norsk. geol. Unders.*
N. J.	*N. Jb. Min. Geol. Paläont.*
N. Ph. J.	*Edinb. New. Phil. J.*
Nat.	*Naturen*
Nat. G. M.	*Nat. geogr. Mag.*

Nat. Mus.	*Natur u. Mus.*
Nat. Volk	*Natur u. Volk*
Njahrsbl.	*Naturforsch. Ges. Zurich, Neujahrsbl.*
Nschr. G. Gött.	*Nachr. Ges. Wiss. Göttingen*
Nw.	*Naturwissenschaften*
Nw. Wschr.	*Naturw. Wschr.*
N.Y. Mus. B.	*Bull. N.Y. St. Mus.*
Nyt. M.	*Nyt. Mag. Naturv.*
Ö. Vet. Ak. F.	*Öfvers. Vetensk. Akad. Förh. Stockh.*
Ostalp. Fst.	*Ostalpine Formenstudien*
P. A. A.	*Proc. Amer. Ass. Adv. Sci.*
P. Austr. A.	*Proc. Austral. Ass. Adv. Sci.*
P. G. A.	*Proc. Geol. Ass.*
P. Lpl. G. S.	*Proc. Lpool. geol. Soc.*
P. M.	*Petermanns Mitt.*
P. P. S.	*Proc. prehist. Soc.* (earlier, *Proc. Prehist. Soc. E. Anglia*)
P. Phys. S.	*Proc. R. Phys. S. Edinb.*
P. R. D. S.	*Proc. R. Dubl. Soc.*
P. R. G. S.	*Proc. R. Geogr. Soc. Lond.*
P. R. I. A.	*Proc. R. Irish Acad.*
P. R. S.	*Proc. roy. Soc.*
P. R. S. E.	*Proc. roy. Soc. Edinb.*
P. Roch. Ac.	*Proc. Rochester Acad. Sci.*
P. Spel. S.	*Proc. Speleol. Soc. Bristol*
P. Y. G. S.	*Proc. Yorks. geol. (polyt.) Soc.*
Pal.	*Palaeontographica*
Pal. Z.	*Palaeont. Z.*
Ph. M.	*Phil. Mag.*
Ph. T.	*Phil. Trans.*
Pr. Z.	*Prähist. Z.*
Q.	*Quartär*
Q. J.	*Quart. J. geol. Soc. Lond.*
Q. J. R. M. S.	*Quart. J. R. met. Soc.*
Qper.	*Quartärperiode*
R. G.	*Rev. Géogr. (annu.) Paris*
R. Hydr.	*Int. Rev. Hydrbiol.*
Rp. B.A.	*Rep. Brit. Ass.*
S. G. F.	*Samml. geol. Führ.*
S. G. M.	*Scot. geogr. Mag. Edinb.*
S. G. U.	*Sverig. geol. Unders. Afh.*
SB. Ak. Wien	*SB. Akad. Wien, math. naturw. Kl.*
SB. bayer. Ak.	*SB. bayer. Akad. Wiss.*
SB. heidelb. Ak.	*SB. heidelb. Akad. Wiss. math. naturw. Kl.*
SB. preuss. Ak.	*SB. preuss. Ak. Wiss. math. phys. Kl.*
SB. sächs. Ak.	*SB. sächs. Ak. Wiss. math. phys. Kl.*
Sc.	*Science (new series)*
Sc. Pr.	*Sci. Progr. Twent. Cent.*
Sc. P. R. D. S.	*Sci. Proc. R. Dubl. Soc.*
Sc. T. R. D. S.	*Sc. Trans. R. Dublin Soc.*
Schr. phys. ök. G.	*Schr. phys.-ökon. Ges. Königs.*
Senckenb. Abh.	*Abh. Senckenb. naturf. G.*
Senckenb. Ber.	*Ber. Senckenb. naturf. G.*
Sm. Ctr.	*Smithson. Contr. Knowl.*
Sm. Misc. C.	*Smithson. Misc. Coll.*
Sm. I. Rp.	*Smithson. Inst. Rep.*
Sv. Mem.	*Mem. Geol. Surv. U.K.*

Svalb. M.	*Svalb. og Ishavet Medd.*
Svalb. Skr.	*Skr. Svalb. og Ishavet.*
T. E. G. S.	*Trans. Edinb. geol. Soc.*
T. G. S. Glasg.	*Trans. geol. Soc. Glasg.*
T. N. Z. I.	*Trans. Proc. New Zeal. Inst.*
T. R. I. A.	*Trans. R. Irish Acad.*
T. R. S. Can.	*Trans. roy. Soc. Can.*
T. R. S. E.	*Trans. roy. Soc. Edinb.*
T. R. S. N. Z.	*Trans. roy. Soc. New Zeal.*
Tijds.	*Tijdschr. ned. aardrijksk. Genoot*
Ups. B.	*Bull. geol. Instn. Univ. Upsala*
U.S. G. S. B.	*Bull. U.S. geol. Surv.*
U.S. G. S. M.	*Monogr. U.S. geol. Surv.*
U.S. G. S. PP.	*Prof. Pap. U.S. geol. Surv.*
U.S. G. S. WSP.	*Wat.-Supp (Irrig.) Pap. Wash.*
V. M. N. F.	*Vidensk. Medd. naturh. Foren. Kbh.*
Ver. I. Meeresk.	*Veröff. Inst. Meeresk. Univ. Berl.*
Ver. Rübel I.	*Veröff. geobot. Ges. Rübel*
Vet. Ak. H.	*K. svenska Vetensk. Akad. Handl.*
Vh. Ak. Wet.	*Verh. Akad. Wet. Amst.*
Vh. G. E.	*Verh. Ges. Erdk. Berl.*
Vh. Gen. Ned. Kol.	*Verh. geol-mijnb. Genoot. Ned. Kol.*
Vh. GTag.	*Verh. dtsch. Geogr. Tag.*
Vh. RA.	*Verh. geol. ReichsAnst. (StAnst.) Wien*
Vid. Selsk. F.	*Forh. Vidensk. Selsk. Krist.*
Vid. Selsk. Skr.	*Skr. norske Vidensk. Akad.*
Vjschr.	*Vjschr. naturf. Ges. Zürich*
Württ. Jh.	*Jh. Ver. vaterl. Naturk. Württemb.*
Y.	*Ymer.*
Z. AV.	*Z. dtsch. öst. Alpenver.*
Z. D. G. G.	*Z. dtsch. geol. Ges.*
Z. G. E.	*Z. Ges. Erdk. Berl.*
Z. G. E. Lpz.	*Z. Ges. Erdk. Lpz.*
Z. Gf.	*Z. Geschiebeforsch.* (*u. Flachlandsgeologie* after 1935)
Z. Gl.	*Z. Gletscherk.*
Z. Gl. Gl.	*Z. Gletscherk. u. Glacialgeol.*
Z. Gm.	*Z. Geomorph.*
Z. Nw.	*Z. Naturw.*
Z. pr. G.	*Z. pract. Geol.*
Zoogr.	*Zoogeographica*

To shorten the references still further, the following types have been uniformly adopted: Series (n.s., N.F., etc.) are in Roman type; part, number, Teil, fascicule are in brackets; volume, year and page (the page in the vast majority of cases refers to the exact page where the statement referred to is made) are in ordinary type and in this order. Where, as in the case for example of the *Neues Jahrbuch für Min.*, etc., more than one volume is issued per annum, the number of the volume is given in Roman numerals after the year, e.g. *N. J.* 1900, II, 89.

Where only a number in italics is given, usually followed by a page number, the reference is to a publication listed in the Bibliography, Vol. II, pp. 1556–1601. When this publication has more than one volume, the number is put in Roman numerals—*1485*, III, 289 is H. B. de Saussure, *Voyages dans les Alpes*, vol. 3, p. 289. When it has more than one edition the number of the edition is given in brackets—*591* (3) is J. Geikie, *Great Ice Age*, Ed. 3.

Titles of papers in the references at the ends of the chapters have perforce had to be omitted. Useful lists will be found in *67, 512, 748, 1763, 1821, 1861, 1862* (3). Titles of American papers are given in *U.S. G. S. B.* 746 (1785–1918), 823 (1919–28), 858, 949, 952, 958, 968, 977; for Switzerland in *Btr. G. K. Schw.* 29 (1770–1900), 40 (1900–10), 56 (1910–20), 73 (1921–30); for Germany in *Geol. Literatur Deutschlands* (publ. by Geol. Landesanstalt, 1897→); for Finland in *Finl. B.* 108 (1555–1933); for the Netherlands in J. F. Steenhuis, *M. G. Dienst.* A (4), 1934; for Sweden in the annual volumes of *Geol. Fören. Stockh. Förh.*; and for the world (outside America) in the bibliography published annually by the Geological Society of America (1933→); *Bibliographie Géographique Internationale, Paris,* (1894→); *Scientific Papers* (1800–1900), compiled and published by the Royal Society of London—18 vols.; *International Catalogue of Scientific Literature* (1901–15), 1903–20; *Geological Literature, Geological Society of London* (1894–1934). Titles of early publications are given in L. J. R. Agassiz, *Bibliographia zoologiae et geologiae*, Ray Society, 1848–54, 4 vols., and recent arctic papers in *Arctic Bibliography* (1947→) published by U.S. Govt. Printing office.

TEXT ILLUSTRATIONS AND THEIR SOURCES

(For italicised numbers, see pp. 1556–1601; for contractions, see pp. xxiii-xxviii)

PLATES

PART I
GLACIOLOGY

(*A*) *LAND-ICE*

Chapter I

SNOW AND ITS DISTRIBUTION

Distribution. The moisture of atmospheric condensation may be suspended as cloud, mist or fog, deposited as dew, hoar-frost or rime, or precipitated as rain, hail or snow. The regions which receive snow are much less extensive than those which are liable to frost.[1]

North and east Europe (Poland and north Russia) have a regular winter snow-cover. In the centre of the continent, the cover in the lowlands is intermittent, in the west occasional, and in the extreme south quite rare. Asia's heaviest snowfalls tend to occur along the periphery from Kamchatka and north Japan to east China. In Africa, snow collects on the Atlas Range in the north, on the great equatorial peaks in the east, and in the south on the Drakensberg scarp and High Veld, on the Basutoland highlands, and occasionally on Table Mountain. North America, the most snowy of the continents, has heavy snowfall in the west from Alaska to California and on the eastern highlands along the great cyclone track as far south as sub-tropical Florida. South America's snow is mainly limited to the Cordilleras. In Australasia, snow falls in Victoria on high ground and sometimes sweeps the Canterbury Plain and southernmost South Island of New Zealand.

Thus snow falls mainly in higher and middle latitudes. In the latter, A. Penck's[2] "seminival" zone of mixed precipitation, the percentage of snow varies considerably; it increases generally towards the poles where snow predominates and diminishes towards the tropics where, save on the highest peaks, snow is unknown.

The extreme equatorial limit of snow is uncertain but in the northern hemisphere[3] probably runs in the region of the Parallel of 30°, passing through the Sahara to embrace Algiers, Tripoli and the whole of the Mediterranean—this sea also freezes occasionally in some places,[4] e.g. in the Adriatic (in the lagoons of Venice and in the Dalmatian bays), in the Aegean Sea, in the Gulf of Salonica, and near Livorno in the Ligurian Sea. From Cairo it crosses the mouths of the Euphrates and Yangtse-kiang to Canton (23° 12′ N. Lat.) and Hainan[5] (18° 45′ N. Lat.) and in the New World passes through the mouth of the Mississippi. In the southern hemisphere, snow has fallen in Sydney (35° 52′ S. Lat.) and in South America as far north as the Tropic of Capricorn.

The summer limit is of course much nearer the poles; except in Norway, it coincides almost with the Arctic Circle and in the southern hemisphere with the 50th Parallel.[6]

The percentage of the annual precipitation that falls as snow, as demonstrated for example for North America[7] (fig. 1), increases in general with latitude. Penck's "full-nival zone", where precipitation is entirely crystalline, is restricted to the Antarctic south of latitude 75° or 77°[8] and to Greenland at heights above 800 m, though even here rain may occasionally occur.[9] Elsewhere in the Arctic the precipitation is mixed.[10]

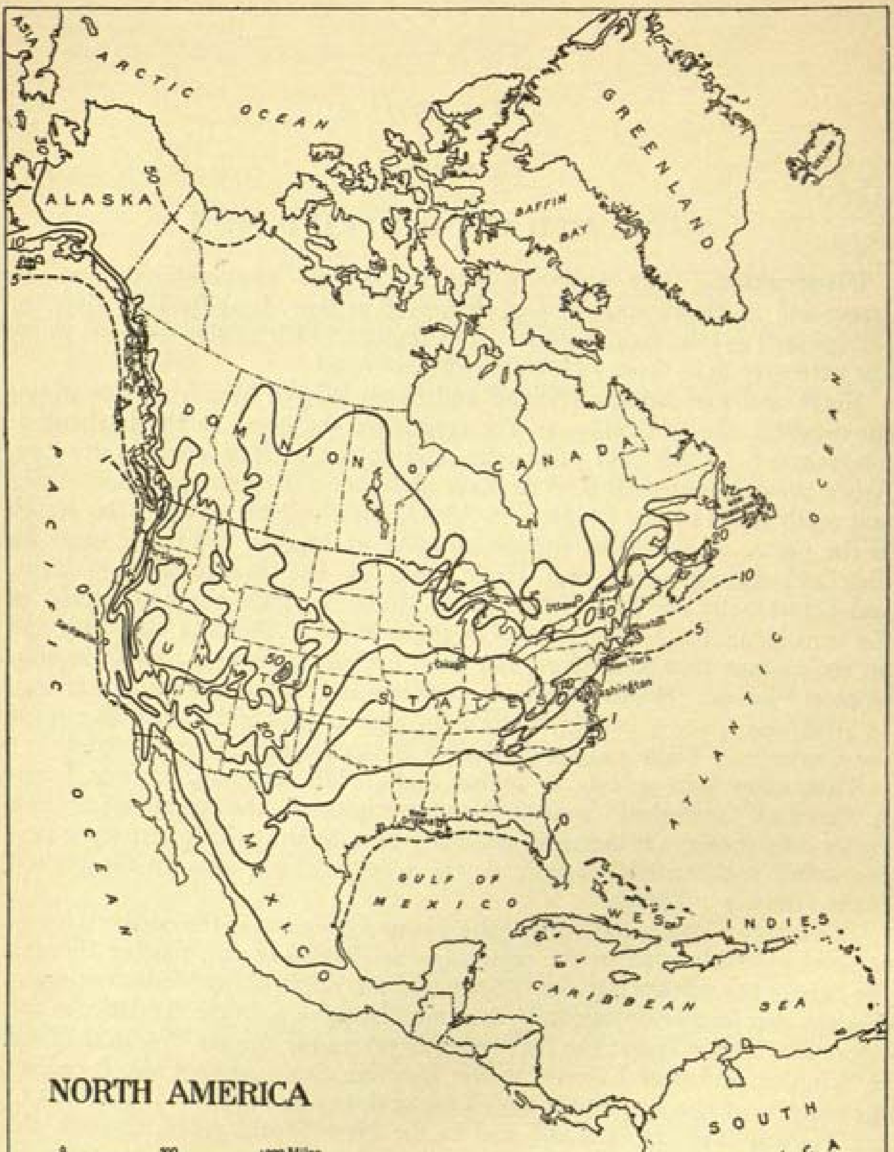

FIG. 1.—Map showing the percentage of the annual precipitation which falls as snow in North America. C. F. Brooks and R. G. Stone, *282*, p. 88.

The ratio of snow to total precipitation increases too with the temperature's fall with altitude.[11] The rate in the Swiss Alps[12] approximates to 3% for every 100 m: the ratio is 10% at 450 m, 28% at 1000 m, 40% at 1560 m, 54% at 2000 m and 81% at 3000 m. Comparable figures for the Eastern Alps[13] are 46% at 1500 m, 59% at 2000 m, 72% at 2500 m, 85% at 3000 m, 93% at 3100 m and *c.* 100% at 3600 m.

Snow and ice conditions have been examined in several countries, e.g.

Finland,[14] and maps showing the mean annual snowfall have been published for north Sweden[15] and North America,[16] including Labrador–Ungava.[17] In general, however, such snow-maps have yet to be constructed.

The number of days in the year when snow lies on the ground, which is so important for both air and soil temperatures and for climate in general[18]—the snow protects the vegetation and ground from frost and by radiation increases the cold of nights—, is shown by equinival or equiglacial lines.[19] Such maps have been made for the British Isles,[20] Tyrol,[21] Finland[22] and for Sweden[23] where the number of days is 60 in the south and 240 in the north. The duration of the snow-cover in Norway is over six months in Finnmark, in the interior of Nordland and on the high fjeld, and sinks to *c.* 20 days on the west coast. In middle Europe,[24] it increases eastwards from the oceanic west—Heligoland and Paris 12, Leipzig 45, Warsaw 60, Riga 100, East Prussia 112,

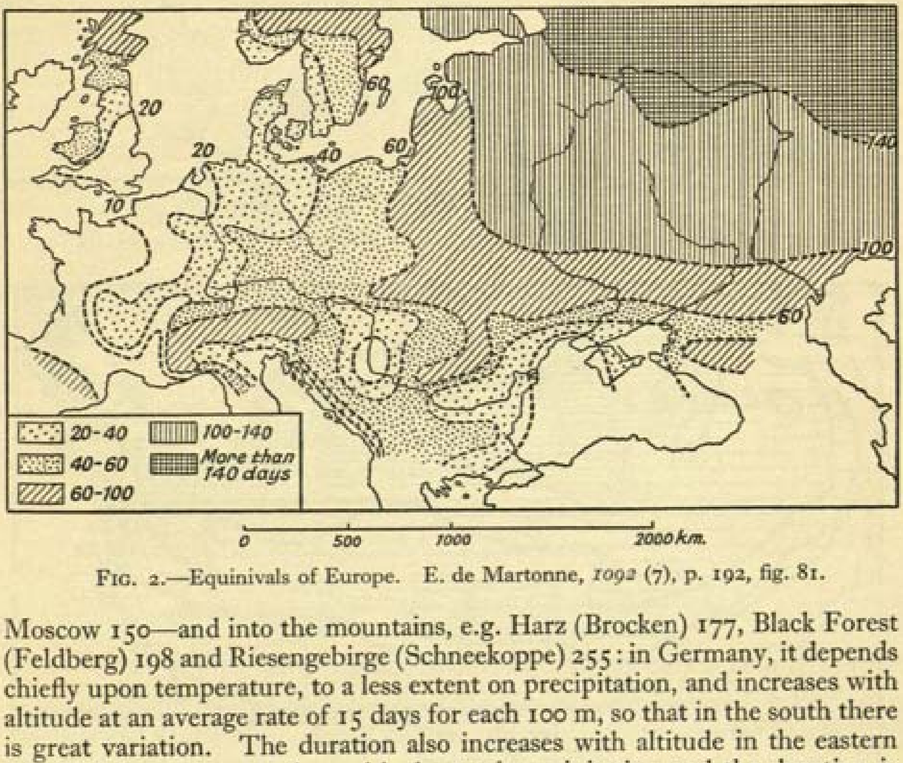

Fig. 2.—Equinivals of Europe. E. de Martonne, *1092* (7), p. 192, fig. 81.

Moscow 150—and into the mountains, e.g. Harz (Brocken) 177, Black Forest (Feldberg) 198 and Riesengebirge (Schneekoppe) 255: in Germany, it depends chiefly upon temperature, to a less extent on precipitation, and increases with altitude at an average rate of 15 days for each 100 m, so that in the south there is great variation. The duration also increases with altitude in the eastern Alps[25] (where the rate rises with the total precipitation and the duration is very great above 2500 m), and in the Austrian Alps[26] where the cover ranges from 38 days at 200 m to 218 days at 2000 m. The duration in different years fluctuates greatly at low levels (32% at 400 m) but is fairly constant above 1200 m (less than 10%). Figure 2 depicts the equinivals of Europe,[27] and fig. 3 those for the United States.[28] Figure 4 gives the duration of the snow-cover in the United States in days per average year.

Character. Although snow surveying was born in Europe towards the end of the 19th century and has important practical applications, the study of the properties of newly fallen snow and the changes this undergoes under different weather conditions is still only in its infancy[29]—the advance of

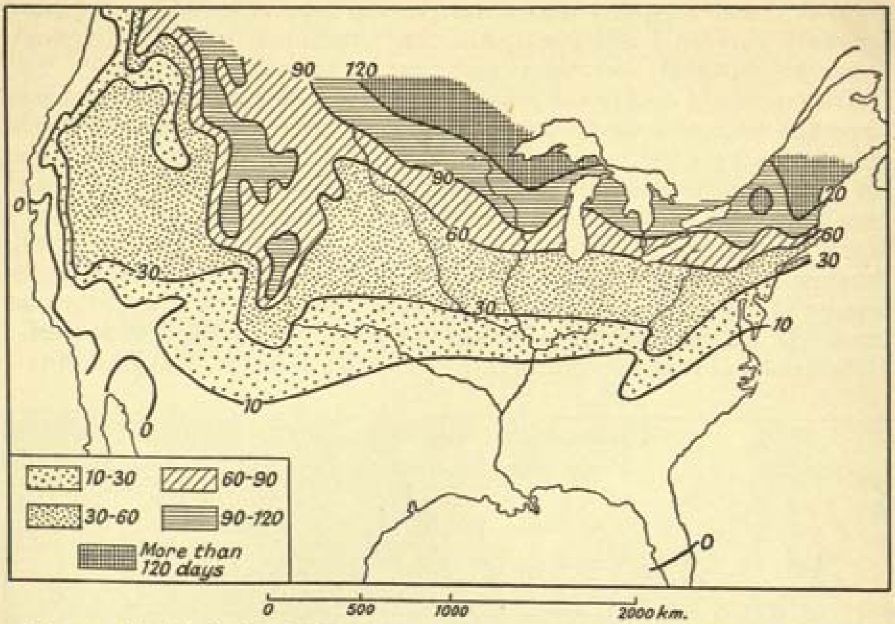

FIG. 3.—Equinivals of the United States. E. de Martonne, *1092* (7), p. 193, fig. 82.

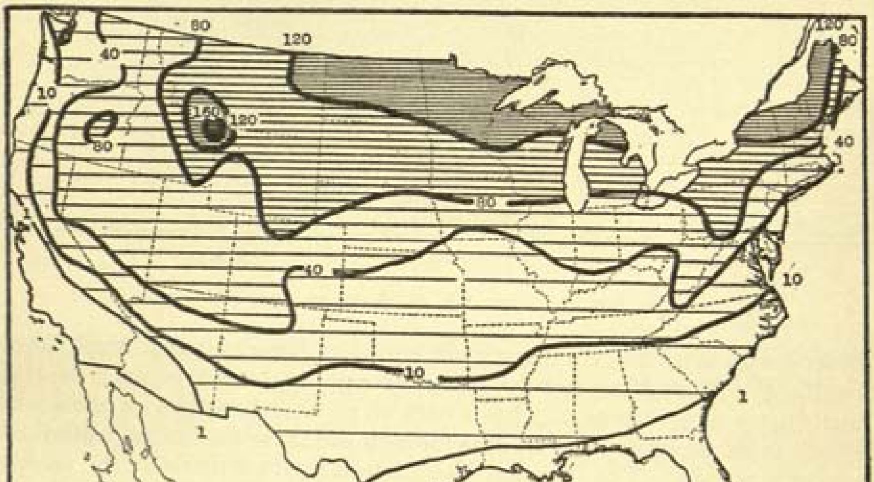

FIG. 4.—Duration of snow-cover in the United States: days per average year. S. S. Visher, *G. S. A. B.* 56, 1945, p. 725, fig. 15.

highways, railways and winter sports into snowy regions has created a demand for the control of avalanches, so that much new knowledge has been acquired of the mechanical properties, structure and metamorphism of different types of snow[30] and of ice and glacier mechanics.[31]

The snow, which chemically and mineralogically is indistinguishable from ice, falls as picturesque needles or flakes[32] (fig. 5)[33]: experiment and the electron microscope have helped to elucidate their growth.[34] The patterns

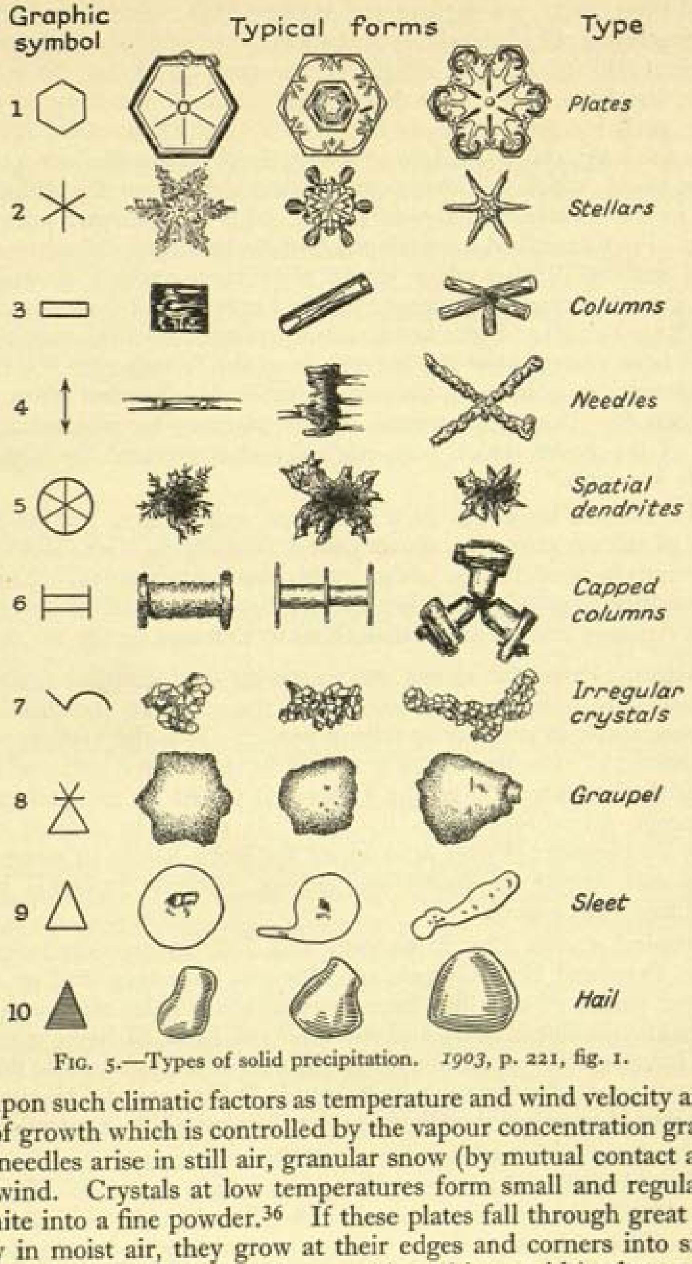

FIG. 5.—Types of solid precipitation. *1903*, p. 221, fig. 1.

depend upon such climatic factors as temperature and wind velocity and upon the rate of growth which is controlled by the vapour concentration gradient.[35] Stars or needles arise in still air, granular snow (by mutual contact and friction) in wind. Crystals at low temperatures form small and regular plates which unite into a fine powder.[36] If these plates fall through great heights, especially in moist air, they grow at their edges and corners into six-rayed stars (less commonly prisms with pyramids or bipyramids) whose rays may add further branches. Thus in general needles or rods, united into bundles or fascicular crystals, occur above the snowline, and stars or needles, coalescing into aggregates or flakes are found near this line. Fern-leaved shapes grow from rapid labile crystallisation, good crystals and prisms from slow metastable crystallisation. Condensation on a glacier yields crystals that differ from those of snowfall.[37]

Polar snow-wastes resemble subtropical deserts not only in their strong

solar radiation, excessive dryness and strong winds, but in their erosive and cumulative forms, of which the simplest are the ripples whose shape depends upon wind velocity and the weight of the snow particles. Snow dunes[38] comprise barchans, free dunes and obstacle dunes heaped up in the lee of rocks or such ice-projections as bergs or a glacier's terminal face. They embrace the long, sinuous ridges or longitudinal dunes, the *sastrugi*[39] (Russ. *zastruga*, sing.), which are long, narrow linear dunes, usually about 2 ft high, aligned in the direction of the wind, and with their steeper ends pointing upwind. First described from Siberia,[40] these trend parallel with the wind[41] ("mixed sastrugi" arise when winds blow from various quarters[42]) and abound in wide expanses of snow,[43] as in Lapland and Greenland and over most of Ross Barrier. Some are erosional, others constructional, though the term has been restricted to the latter[44] or to the former type.[45] Compared with their analogues in sand, they are generally less flat and grow and move less quickly.[46] The snow crystals freeze together by regelation, and the density of the cover which was originally characterised by high porosity gradually increases.[47]

On the steeper lee edges in a landscape, e.g. ridges, plateau scarps or margins of stream channels, snows collect as cornices (Ger. *Wächte*) which may be temporary or may be winter or permanent structures. The cornices grow according to aerodynamic laws, which have been recently investigated,[48] until overloading and melting cause them to sink and finally break off.

Snowline. Perennial snows on temperate and tropical peaks and in polar wastes are bounded by a lower line, the so-called snowline, where as much snow melts in summer as falls in winter. It is the highest position of the "temporary" snowline which generally lies at a lower level and fluctuates widely with the seasons[49]: on the Pasterze it varied by as much as 600 m. The concept, introduced by L. Bouguer[50] in 1736, was defined more precisely by de Saussure[51] who determined the line's height in many places in the Alps and sought the controlling factors. Several writers[52] have dealt with the historical aspect.

On tropical domes like Ruwenzori and Kilimanjaro, on mountains in Ecuador, Peru and New Guinea, on Mexican volcanoes, and on the outer Himalayan peaks, all of which have more or less regular outlines and a uniform climate, the line is sharp and regular[53] (pl. Iᴀ, p. 8), being generally unaffected by orographic patches and but little modified by seasonal fluctuations or by aspect because of the sun's high altitude.[54] It was indeed in these regions of simplicity of snowline that Bouguer first recognised the controlling factors.

The line loses its distinctness in high latitudes where it expands into a zone of indefinite breadth which cannot be fixed within 400 m.[55] Thus in parts of Greenland[56] the zone is 20–50 km broad and has a vertical interval of 1000 m. This is because the sun's rays have a very low angle, the relief at these relatively low levels is varied, and the ice-sheets incline only slightly.[57] The wide yearly oscillations of the limits[58] (cf. the mean and maximum snow altitudes in Baltoscandia during the winter months[59]) which find their climax on isolated mountains[60] add to the indefiniteness: on the Pasterze[61] (see above), snow covers the tongue from November on and melts only in June, while in Steiermark[62] the temporary snowline stands as follows: March, 700–900 m; May, 1300–1800 m; July, 2400–2600 m; September, 2400–2800 m; October, 1700–1900 m; November, 1000–1300 m (cf. p. 14).

Controlling factors. The snowline coincides with neither contour nor isotherm but zigzags acutely up and down according to summer heat, radiation, reflection, precipitation, humidity, cloudiness, force and direction of the prevalent wind, avalanches, proximity to glaciers, the nature of soil and rock, rock-structure and slope aspects.[63] G. Wahlenberg[64] (1813) emphasised the meteorological factors, and A. v. Buch[65] attempted to ascertain those, both meteorological and orographic, which govern the Norwegian snowline. A. v. Humboldt[66] recognised the factors still more clearly.

The snowline is obviously influenced by the summer warmth or the mid-day temperature of the hottest month, e.g. in the Swiss Alps.[67] The relationship is brought out by the course of the snowline through the latitudes and its connexion with the temperature curve[68] (fig. 6): in both lines, the equatorial depression lies north of the equator and the southern peak is higher than the northern one. The snowline was therefore early identified with the summer isotherm of 0°C.[69] Yet this is manifestly erroneous[70] since the line is not solely a function of temperature: precipitation is also of cardinal

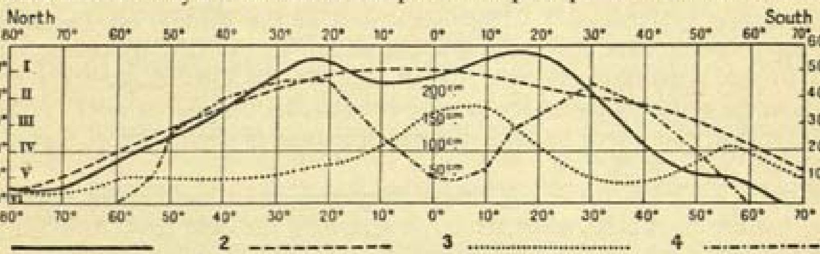

FIG. 6.—Mean altitude of the snowline (1), according to latitude compared with temperature (2), precipitation (3), and the index aridity (4). E. de Martonne, *1092* (7), p. 321, fig. 117.

importance. Although the line sometimes coincides with the summer isotherm of 0°C, as in parts of Scandinavia,[71] or with the zero isotherm of the warmest days, as in west Greenland,[72] this usually is not so. Thus in the Jotunheim area and in Norway generally the temperature at the snowline is higher in the wetter west than in the drier east.[73] Diminished precipitation explains the slight fall or even rise in the snowline northwards through west Greenland[74] and the slow motion and gentler slope of the ice-sheet in north Greenland.

In average climates between 35° and 70° N. Lat. the line coincides with the isotherm of 4°C±3°C of the warmest month or −4°C±2°C of the year.[75] V. Paschinger[76] states that it oscillates between 10°C and −10°C and lies in the Alps at −4°C and in central Asia between −6°C and −8°C. Near Mount St. Elias in Alaska it is 10°C, in the Sierra Nevadas of California 8·5°C, in the Andes near Junin −7°C and on Kilimanjaro −4°C. H. W:son Åhlmann[77] placed it in the Alps and Norway at the summer isotherm of 2°C and drew a map showing its relation in the Alps to the summer isotherms. In Iceland it ranges from −0·7°C in the east-central region to 4·3°C on the north-west coast.[78]

Latitude is often not permitted to exert its full effect for snowfall offsets its influence. Thus Alaskan glaciers increase in number and size and their snouts decrease in altitude towards the centre of the great semicircle surrounding the head of the Gulf of Alaska[79]; the snowline is at a low altitude in north Iceland and Patagonia[80] and is locally depressed by bodies of water,[81] such as Lake Titicaca, the Black Sea and Caspian Sea. It is also lower on

the western edge of the Pamirs[82] and, despite their higher temperatures, on the southern slopes of the Caucasus[83] and Himalayas[84] (the difference ranges up to 2000 m). On the south side of the Himalayas it coincides with the mean annual isotherm of 0·5°C to −1°C and on the Tibetan side with one of −4°C to −5°C. The effect is also seen in the dimple in the latitude curve near the equator (see p. 9)—the minimum is *c.* 10° N. Lat.—and in the discontinuous ice-cover of both to-day and the Ice Age in northernmost lands.

Thus temperature and precipitation vary regionally in importance.[85] In general, precipitation is dominant where it is heavy, as in lower latitudes, eastern Alps, Caucasus, peripheral Asia, and west America, or if temperatures are low, as in north-west Greenland whose snowline is high because the precipitation is unusually low. Temperature, on the other hand, is supreme if the snowfall is light, as in the western Alps, central Spitsbergen, and central Asia, including the Russian Altai[86] whose snowline is at 2500 m on the drier north and at 3000 m on the moister south, and in the Himalayas where the snow falls in the hot season, the loss by ablation is large and the high, steep walls favour avalanches.[87] The oscillations of the firnline on Vatnajökull during 37 years were determined to the extent of 75–80% by precipitation and to the remaining 20–25% by ablation.[88] In Greenland[89] the coast lands in the south and west are ice-free because the summer temperatures are high, and in the north because the precipitation is slight (see p. 643): the optimum is in 75° N. Lat.

Wind is a third factor.[90] It nourishes glaciers in Lapland below the snowline,[91] while elsewhere, when strong, it dislodges the light snow to lower and warmer levels where the snow melts or, in the cold polar regions, e.g. in Greenland, Norway and the Ural Mountains, builds "snowdrift glaciers" far below the general snow limit.[92] In the Antarctic, winds are the chief means by which the snow is removed, exceeding possibly the glaciers themselves[93]: in summer, they cause ablation at the foot of the mountains about Ross Barrier.[94] Elsewhere they regulate the distribution of snow and orientate the glaciers[95]—it is estimated that *c.* 20,750 cu. m were swept over one Alpine pass in one winter (1935–6).[96] Strong winds, which blow the snow into the sea, are forcing an extremely rapid retreat in Novaya Zemlya.[97] On the even plateaux of north Greenland, e.g. on the northern coast of Peary Land and of Jameson Land, as well as on Bear Island and in places in Spitsbergen, e.g. Reindeer Valley, the Sierra Nevadas of California (including Mount Whitney, the highest summit in continental United States), the South American Cordilleras, and in Kerguelen, they carry the snow away so that glaciers are unable to form,[98] unless sheltered by a well-marked topography. In east Greenland,[99] slopes and plateaux are swept clean up to altitudes of 1500 m, though the mean annual temperature is less than −15°C. Even glacier oscillations, as in north Greenland, are related to the winds.[100] Moreover, as Enquist[101] showed for Scandinavian and other mountains, snowfields and glaciers occur mainly in the lee where snow, in contrast to rain, tends to accumulate, a contrast emphasised already by others[102] and established for various countries.[103] The difference in height of the snowline between sun and shadow sides is 200–400 m in the Alps[104] and over 1000 m in the Pamirs.[105]

The amount of snow-drift depends on the consistency of the snow and especially upon the velocity of the wind. Drift begins if the wind force is 6–8 m/sec and increases with probably about the fourth or fifth power of the competent velocity.[106]

Types of snowline. Fixing the snowline's precise position (Wahlenberg's *terminus nivalis*) is made difficult not only by the sun and shadow but by irregularities in the mountain flanks. Instead of a straight line parting continuous snow above from bare rock below, the upper zone has steep or vertical faces free from snow and the zone below has snow-filled hollows of various sizes. The line joining the lowest edges of the perennial snow patches in the *regio subnivalis* is Ratzel's[107] "orographic snowline" (Wahlenberg's *terminus subnivalis*): avalanches which descend still lower (avalanche glaciers, facing north, may be wholly below the snowline) and snow patches at the foot of screes are ignored in determining it.[108]

Snow nestles below the snowline[109] in sheltered ravines, clefts or chasms, in the angle at the confluence of valleys, in gullies notching cirque walls, and at the head and foot of screes. These *bas-névés*[110] of Charpentier and the brothers Schlagintweit constitute Ratzel's *Firnflächenlandschaft* and are bounded below by his *Firnfleckenlinie.*[111] If the relief is suitable they may fall into a double zone, as in parts of the Limestone Alps.[112]

The climatic or regional snowline, defined by some as the mean snowline on the shadier side of mountains, is generally regarded[113] (in approximate accord with Wahlenberg's original definition) as the snowline on horizontal or neutral surfaces which are not interfered with orographically or influenced by sun or shade, wind or lee, or wet or dry sides. This surface, being ideal or theoretical, is mostly indirectly ascertainable. The simpler the relief and the narrower the zone of detached snow patches, the closer do the orographic and climatic snowlines converge.

Enquist's *Vergletcherungsfläche*[114] (glaciation limit), the inclined plane which parts all summits that nourish glaciers from those that do not, is generally parallel with the climatic snowline but about 100 m higher. E. Brückner[115] regarded the two as identical but H. W:son Åhlmann[116] deems them to be separate and calls the lines joining places where the glaciation limit is at the same altitude *isoglacihypses*. He has made such a map for modern Iceland and shown that the smallest value is 700 m and the highest, viz. 1400 m, is at the extremely arid centre. Somewhat similar is the *Schneegrenzfläche* of H. Louis,[117] defined by him as the plane which, continued over intervening stretches in the air, would have glaciers above it if the land projected high enough: Louis constructed this plane for western U.S.A.

The difficulty of locating the snowline on mountain sides, which caused J. Payer[118] to deny its very existence, led Hugi[119] to substitute for it the firnline on glaciers, this in his opinion being more definite and constant and independent of the slope and of the type of rock and vegetation. Hugi's suggestion was subsequently adopted[120] and is acted upon to-day. Yet the firnline varies from year to year, from month to month and from glacier to glacier.[121] Its course is seldom straight and depends on a glacier's orientation. On north–south glaciers like the Aletsch it runs roughly horizontally and transversely; on east–west glaciers, e.g. the Lower Aar Glacier, it is much higher on the northern side which is exposed to the sun's rays—on one Jämtland glacier the difference was 170 m.

The névé-line is lower than the snowline on adjacent rocks[122] since the conductivity and slope are different. Glaciers likewise cool contiguous rock-surfaces and raise the percentage of precipitation which falls as snow while the cold winds or aerial cascades, of varying but shallow depth, which blow down them[123] depress the firnline as de Saussure[124] noticed in the Alps and

von Buch[125] in Norway. In the Ortler group, the depression is 40 m,[126] in the Oetztal group 200 m,[127] and in Spitsbergen above this figure.[128]

Determination. Several methods, all indirect and practically inapplicable to any but cirque and valley glaciers, have been adopted to fix the snowline. One of the earliest, Höfer's method,[129] which F. Simony and J. Partsch had already used,[130] assumes that the snowline is the arithmetical mean between the altitude of the glacier snout and the average height of the crest above the firn. This method, which has for a corollary that the depression of the snowline at a recent stage is half the altitude between the glacier-end now and at the stage in question,[131] is inaccurate since there is no such simple relationship. On the contrary, a glacier's descent depends upon the size of its firn basin, its aspect and snowfall, the build of its bed, and the amount of its debris-cover[132]—the vertical interval between the snowline and the snout in the present Alps is often 1500–2000 m.[133] The method, moreover, ignores the glacier's shape and is otherwise unsatisfactory.[134] It may only be applied to small glaciers or to those which slope uniformly, e.g. the Vernagtferner, or those in which the upper concave part roughly equals the lower convex part, as on the Pasterzeferner.[135]

The Partsch or "summit method", suggested by de Saussure and Simony[136] and afterwards used for Pleistocene snowlines,[137] assigns limiting values: peaks and ridges which nourish glaciers or névés give an upper one, neighbouring summits without such accumulations give a lower one. Its results are good if many observations are available. While it depends less upon orography than do other methods[138] it is inapplicable to summits which project far above the snowline, and is difficult to apply to plateaux.

Brückner's hypsometric method,[139] based upon observations in the Hohe Tauern, aims not so much at ascertaining the snowline of individual glaciers as that of a whole region. Brückner measured the snow- and ice-covered area of a mountain group, subtracted one-quarter for the glacier tongues, and computed the altitude layer which equalled this in area. The method, which requires a planimeter, assumes therefore that the ratio of firn to tongue is 3:1. But the actual ratio often departs widely from this[140]; in the Alps it is between 1·5:1 and 7·5:1 (average 3:1) and during the Pleistocene was between 1·5:1 and 3:1 and in the Niedere Tauern and Gurktal Alps varied between less than 1:1 and more than 3:1.[141] In the eastern Alps, it is nearly 4:1, in the Oetztal and Hohe Tauern 3·8:1 (see below). On Mount Hood and Mount Adams in the Cascade Range again it differs materially, being 1:3, possibly because their tongues are very broad.[142] The firn is also relatively large in the wide, short glaciers[143] of Iceland and Norway where the ratio is 1:1, and in Franz Josef Land and the Hohe Tauern sometimes almost coincides with the glaciers.[144] In the Karakoram Mountains on the other hand the tongues are relatively long, namely up to 5:1, because surface debris is abundant[145] and the high enclosing walls cast long shadows. The method is therefore quite empirical and is difficult to apply to modern glaciers and even more to their Pleistocene ancestors. Like the previous methods, it exaggerates the height of the snowline, as Brückner himself recognised, since the snow-free areas in the firn basin are not omitted from the calculation. Richter[146] suggested that when using it, choice should be limited to glaciers in comparatively wide and shallow depressions.

Kurowski's method,[147] of which Brückner[148] gave a mathematical proof,

was founded on investigations in the Finsteraarhorn group. It assumes that precipitation and melting are linear functions of altitude, the one increasing, the other decreasing regularly with this: where they are in equilibrium is the glacier's true mean altitude or firnline. Applied to a single glacier, it gives the orographic snowline, applied to a group of glaciers of differing aspects, the climatic snowline. This method, though not without its blemishes,[149] has been found reliable.[150] H. F. Reid,[151] from considerations of lines of flow, showed that although it strictly applied only to glaciers in equilibrium, the error was very small if these were not varying much; Kurowski estimated the error in the Alps at only 20–30 m. The method gives higher values than other methods since precipitation and ablation do not vary as regularly as Kurowski supposed: that this is so has been shown in the Caucasus and French Alps,[152] especially if allowance is not made for the snow-free margins in firn basins. Like Brückner's method, it can only be employed with good contour maps. It is inapplicable to Pleistocene glaciers unless these had a definite "erratic limit". Circumventing this difficulty by substituting for the glacier-surface the drainage basin within the outer moraine[153] has given a method that has proved satisfactory in north Switzerland,[154] except perhaps in valleys which have a very small fall.

Probably one of the best methods, though it is neither very accurate nor available for glaciers of the Ice Age, is Hugi's morphological method.[155] This is founded upon the difference in the contour's shape in the firn and tongue, a difference early recognised[156] in Spitsbergen and in Norway. While the contours bend sharply where the convex tongue meets the valley sides, they are concave in the firn and pass without a break to the bounding walls because the snow accumulates along the base of the cliffs which reflect and radiate little heat. The transition is the firnline. The method seems obviously correct and is found to hold[157], e.g. in the massif des Grandes-Rousse, Savoy Alps and Novaya Zemlya, but is disputed for Jotunheim.[158]

Certain well-known relationships provide auxiliary aids. The floors of the lowest cirques in a mountain group are about the level of the local snowline[159] (see p. 296) while the bergschrund lies above, dirt-bands and surface moraines below it. The last relationship, whose soundness Finsterwalder theoretically demonstrated (see pp. 121, 411), has often been employed[160] (e.g. in the Alps, Savoy, Hohe Tatra, and Sierra da Estrela, Portugal)—the "firn moraines" have no genetic connexion with either true inner or surface moraines.[161] Finally the lines of flow are emergent in the tongue, immergent in the firn, and tangential at the firnline.[162]

The snowline's altitude on plateau or continental glaciers is less readily obtained. It may be fixed by ascertaining the altitude of the marginal moraines, the snow-free areas on the glacier and the rocky surfaces, or the relationship of the surface snow to the ice below,[163] as got by profiles dug in the glacier, e.g. in Greenland, Iceland, Spitsbergen and North-East Land: above the snowline the one grades into the other but in the ablation zone there is a sharp break.

Data. Since Humboldt[164] first gathered together the data of the snowline's altitude, a vast mass of material has been collected which Paschinger[165] sifted and summarised. Unfortunately, the older determinations are often unreliable and inaccurate and the methods employed are rarely stated. Seasonal and yearly fluctuations were usually neglected. The former,

though noted by Venetz as early as 1821, have only recently been thoroughly investigated,[166] as in the Swiss Alps and the Jura Mountains. Their importance may be gauged from data[167] for the Inn valley where the snowline in December–January lay at 600–700 m and in August–September at more than 3000 m, and from the corresponding ones of 725 m (March) and 2397 m (September) in the Säntis and of 2500 m and 2800 m for the Rhône Glacier. The seasonal variation on a glacier of low gradient may be several kilo-

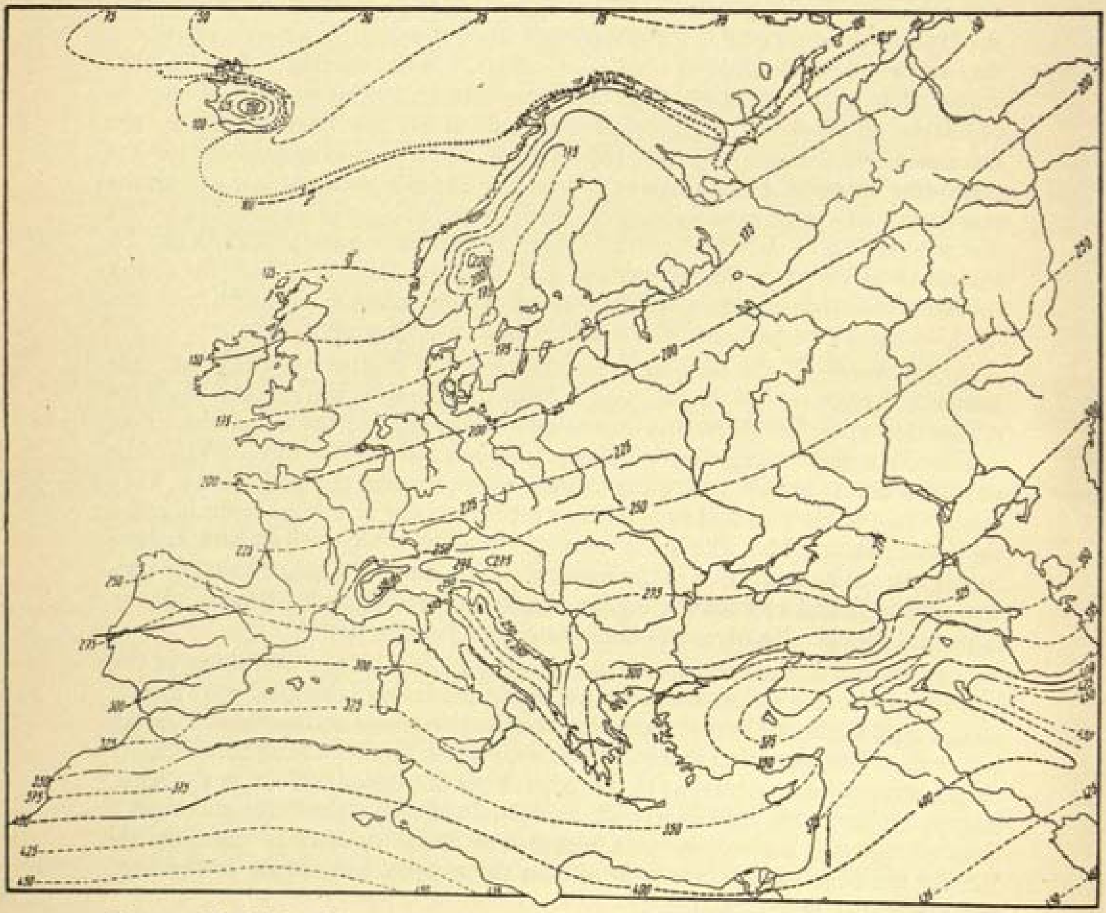

FIG. 7.—The height of the snowline (in decametres) in Europe, Asia Minor and North Africa. Line of crosses = the present July isotherm of 10·5°C at sea-level. J. Büdel, *Nw.* 36, 1949, p. 106, fig. 2.

metres.[168] Hence observations must be spread over many years to get the average altitude, which in the most favourable circumstances can only be expressed in whole numbers and not within 50 m.[169] It is therefore hazardous to assume vertical crustal movements from local discrepancies in the height of the Pleistocene snowline, as has been done in the Balkans.[170]

The Alpine snowline has been studied by many workers,[171] including Richter (eastern Alps), Jegerlehner (Switzerland), Kurowski (Finsteraarhorn group), R. Zeller (Trift region), Paschinger (French Alps), and M. Fritsch (Ortler group). Its course in Europe and the Mediterranean region is shown in fig. 7.[172] In the Alps it is highest on Gran Paradiso (3350 m) and sinks

rapidly westwards (Mont Blanc group, 2800 m), but less rapidly northwards to Monte Rosa (3200 m) and Säntis (2504 m).

The Norwegian snowline[173] (fig. 8),[174] examined by A. M. Hansen[175] and G. Holmsen[176] and more fully by Åhlmann,[177] reaches its highest point just south of Dovrefjeld and lies on Svartisen at 800–1100 m, on Folgefonn at 1450–1500 m or 1300–1400 m, on Jostedalsbrae at 1500–1600 m, and on Jotunfjeld at 1900 m. It falls to below 1000 m in the Lofoten Islands and to below 800 m in parts of Finnmark.

The altitude in the Sierra Nevada is 3200 m.[178] Elsewhere in Europe it is above the highest summits: in the Riesengebirge and Black Forest it may be at 2100–2200 m.[179] Ben Nevis, Braeriach and Ben Macd'hui in Scotland[180] escape glaciation by a narrow margin, for the orographic snowline was probably above Ben Nevis at 1450–1500 m between 1884 and 1903 (see p. 646) where the summer temperature may be 3·6°C: a fall of 1·6°C in the average temperature would initiate "infant" glaciers. Snow-banks,[181] almost permanent, nestle on the north-east face of the mountain (they disappeared in 1933, 1935 and 1945) and in gullies on Braeriach and Ben Macd'hui in the Cairngorm Mountains (they disappeared in September 1933 after persisting, it is believed, since before 1864). These "outliers" of the climatic snowlines are due to drifting snows, helped liberally by avalanches from cornices. Less permanent snows lie in Y Ffos Ddyfn near the summit of Carnedd Llewelyn in Wales.[182] Similar small névés occur on Feldberg in the Black Forest,[183] in the south Carpathians,[184] in the eastern Alps,[185] in Gran Sasso (2914 m; 42° 28′ N.), central Apennines,[186] in the Corral di Veleta, Sierra Nevada,[187] in Corsica (W. Heybrock, 1954) and with 12 or more miniature glaciers in eastward-facing cirques at altitudes of up to 1885 m in the north Urals in 64° 40′–65° 40′ N. Lat.[188] In the mountains about the North Atlantic with a maritime climate the snowline probably lies at least 450 m higher than the exceptional sheltered snow-patches.[189]

The Caucasian snowline,[190] recently examined by several investigators,[191] lies on southern slopes, in the west at 2700 m, in the east at 3250 m, and farther east at 3700 m; and on northern slopes 300–400 m higher. The height is about 3500 m in north-west Persia[192] and in Anatolia[193]: a recent map of H. Louis[194] shows the snowline of Asia Minor rising in concentric ovals from *c.* 2400 m to a maximum of 2800 m or even 3700 m in the mountains (Ala Dag) north of Adana. The altitude in central Tienshan[195] (40°–46° N.) is 3450–4000 m or 3800 m on the north and 4200 m on the south, in east Tienshan[196] 3652 m on the north and 3937 m on the south, and in west Tienshan[197] varies between 3500 m and 3900 m. In the Altai[198] (46°–50° N.) it is 2000–4000 m and in the Himalayas[199] is *c.* 4270 m in the east and 5800 m in the west and *c.* 915 m higher on the Tibetan side. In the Karakoram Mountains and west Himalayas[200] it runs between 4250 m and 6250 m, in the Pamirs[201] at 4000 m in the west and 5200 m in the east, and in Kuenlun[202] at 6000 m (cf. G. Dainelli's map[203] of Karakoram–Himalayan snowline). In Japan,[204] it is slightly above the highest summits.

The snowline on the isolated peaks of equatorial Africa is at *c.* 7625 m[205] and as on other tropical peaks is considerably lower on the western than on the eastern side.[206]

Arctic conditions are more complicated; the frozen ground, cold and warm currents, and drift-ice (notably in coastal Greenland and Franz Josef Land) introduce modifications.[207] The snowline is everywhere above sea-level,

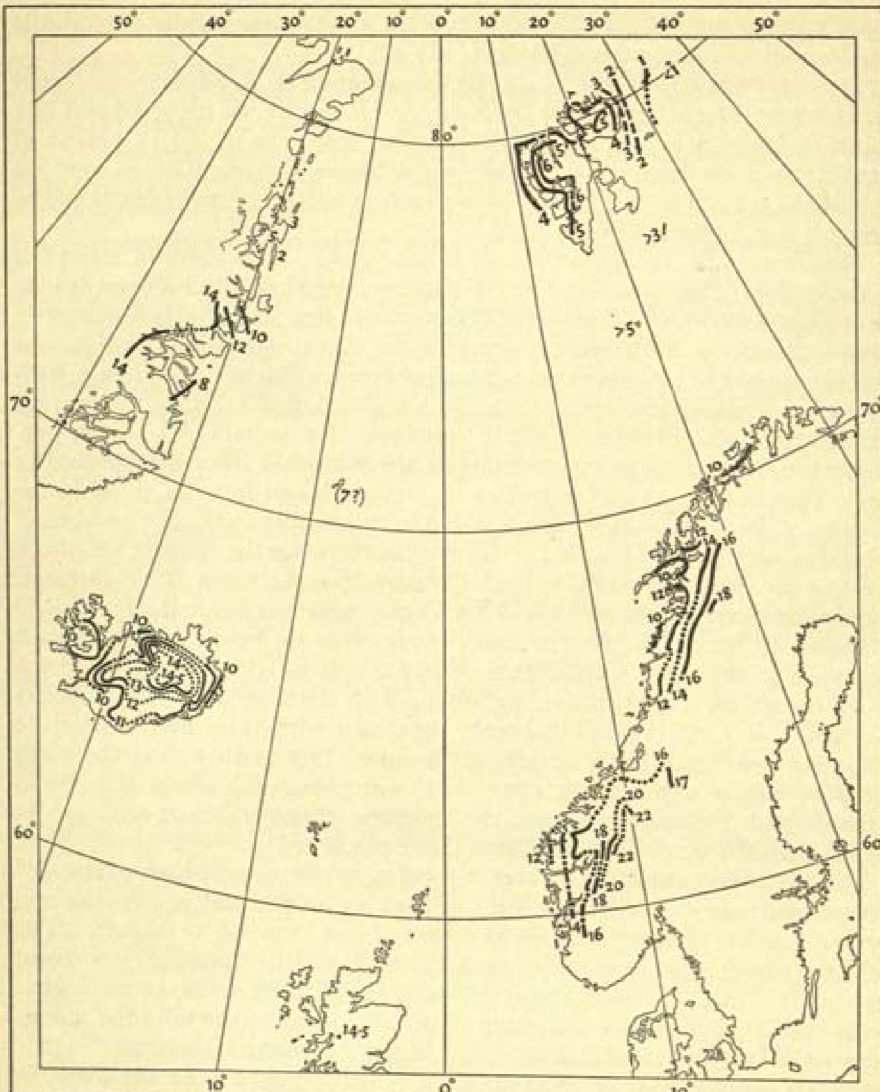

FIG. 8.—The course of the snowline or isoglacihypses (in hundreds of metres) in the lands around the North Atlantic Ocean. H. W: son Ahlmann, *25*, p. 46, fig. 31.

except orographically in a few places in Greenland.[208] Around the Norwegian Sea it rises concentrically with the coast into the surrounding lands [209](fig. 8). In Iceland[210] its altitude is 400–1600 m (Vatnajökull, 1050–1400 m; Hofsjökull, 1300 m). On Bear Island,[211] which has no ice-fields or glaciers, it is at least 180 m or even 550 m; in Spitsbergen,[212] 300, 600 or 1000 m, dropping rapidly eastwards to *c.* 150 m in North-East Land and to *c.* 100 m in White Island, to lower still in Victoria Island, and to below 70 m in northern Severnaya Zemlya. In Jan Mayen[213] it is at *c.* 700 m, in Novaya Zemlya[214] at 450–590 m and in Franz Josef Land at 100–120 m.[215]

The altitude is uncertain in Greenland,[216] whose accumulation area covers 83·5–84% of the whole ice-sheet[217]; in the west, where it is best known,

it is 700–800 m in 69° N. Lat., and 860 m on Nûgssuaq Peninsula, but generally between 64° and 78° is *c.* 1400 m. In south Greenland it is 700–900 m and 300–350 m only near Cape York, where southerly winds carry the snow on to the unprotected coastal strip and incidentally keep the North Water permanently open (see p. 192). On the north coast in Peary Land, where the climate is continental (January mean temperature −32°C, July mean temperature 6°C and annual precipitation 114 mm), it is at 1250 m and in the east at 900–1200 m or even 1500 m or 2000 m.

While its height is relatively low near the coasts and on the adjacent islands (in east Labrador it is lower still[218]), it rises rapidly inland[219] to 700–800 m at the edge of the ice-sheet and to 1450–1530 m in the interior.

The North American snowline[220] (see p. 93) falls generally northwards; it is at 12,000 ft (*c.* 3700 m) in the High Sierra and at 2000 ft (*c.* 610 m) in Mount St. Elias, Alaska, but rises again by 4000 ft (*c.* 1220 m) in 80 miles (*c.* 130 km) to reach 8000–10,000 ft (2440–3050 m) in the Aleutian Mountains. In northern Ellesmere Land it is at 3400 ft (1036 m), in Baffin Land at 5100 ft (1555 m).

The data amassed by R. Hauthal, H. Meyer, W. Sievers and others for South America have been examined by Ratzel[221] and by G. Schwarz[222] who drew a profile of the snowline from South America to Mexico. The following table shows its course through the latitudes:

Latitude	*Altitude*	*Latitude*	*Altitude*
10° N.	4700 m	34° S.	3550 m
10° S.	4900 m	36° S.	2600 m
16°–21° S.	6000 m (W.)	41° S.	1560–1700 m
,,	5400 m (E.)	43° S.	1400 m
24° S.	6200 m	53° S.	1100 m
30°–32° S.	3500 m	54° S.	950 m

It is highest between 16° and 18° S.,[223] is at 1000 m in 47° S. Lat., and leaves South America at 500 m[224]: the greater glacierisation changes from the east to the west side of the Andes in 38° S. Lat.[225] By contrast, no Australian mountain attains the snowline.

The Antarctic snowline runs at or below sea-level[226] and outside the coast; for on Ross Barrier and in Queen Mary Land, Graham Land and other places along the continental margin the snowfall exceeds the loss by wind, melting and evaporation and (contrary to other opinion[227]) the land is above the zone of ablation[228]: equilibrium is maintained by the snow broom and by calving. The ice is covered with névé to its termination; a zone of melting is lacking; and dust-wells occur on horizons at different depths.[229] The alimentation of the pack-ice surrounding the continent is further proof. Nevertheless, just as in the Arctic, there are departures from the general rule. The orographic snowline, as ice-free areas and surface-moraines show, lies exceptionally above sea-level,[230] as near Gaussberg, near Cape Adare, in Graham Land, and in West Antarctica, where the action of the wind, the relief, or the nature of the rocks are unfavourable to the accumulation of snow. Rain is only known in Graham Land where the land projects farthest north.

The contrast between the Arctic and the Antarctic is very striking. In Patagonia and New Zealand, the glaciers are larger than in similar latitudes in the northern hemisphere: Franz Josef Glacier reaches to about 215 m above sea-level in 43° S. Lat. Kerguelen is glacierised in 50° S. Lat., corresponding to southernmost England in the northern hemisphere, and glaciers

reach the sea in Heard Island (53° S.), South Georgia (54° S.) and Bouvet Island (54° 26′ S.). The area above the snowline in the Antarctic is probably *c.* 21 million sq. km, in the Arctic only 800,000 sq. km,[231] while the ice of the southern hemisphere has an area of 13 million sq. km and a volume of 22·18 cu. km, the corresponding figures in the northern hemisphere being 2·1 million sq. km and 2·7 cu. km[232].

Although the snowline occasionally coincides with the sea-level in the Antarctic islands,[233] e.g. in Heard Island and the South Orkneys, it is generally well above that level as will be gathered from its various altitudes[234]: South Georgia, 300–650 m; Kerguelen, 600 m; Possession Island, 700–800 m; Tierra del Fuego, 1050 m; Crozet Island, 1000 m; and Bouvet, about 200 m.

Laws. The foregoing figures, representative but seemingly unrelated, conform on analysis to a few simple laws. According to the first, the snowline declines from the tropics to the poles (fig. 6), as Bouguer[235] recognised, the gradient being steeper in the southern hemisphere: in New Guinea the snowline is at 4650 m, in the North Island of New Zealand about 2400 m and at the southern extremity of the South Island *c.* 1830 m.[236] It falls *c.* 1200 m between 24° and 41° S. Lat. in the Andes[237] and in South Georgia is lower than the treeline of Tierra del Fuego. This behaviour is illustrated in Norway[238]: 63° N., 1355 m; 66° N., 1245 m; 68° 30′ N., 885 m, and by the following figures for the Apennines[239]:

M. Pollino	*M. Cervali*	*M. Miletto*	*M. Terminillo*	*M. Pisanino*	*M. Beigua*
39° 54·5′	40° 20′	41° 27′	42° 28′	48° 8′	44° 26′
2000 m	1840 m	1750–1800 m	1700–1750 m	1300–1400 m	1000 m

The line is very high near the equator, as the figure of 5750 m on Kilimanjaro exemplifies[240] (H. Meyer[241] summarised the tropical data). Small glaciers occur in America on Sierra Nevada de Santa Marta (11° N.), Sierra Nevada de Cocuy (6° N.), Nevada del Tolima (5° N.), Nevada del Huila (3° N.), Sierra Nevada de los Coconucos (2° N.), Quito (0°), Palmira (2° S.), Páramo de Almorzadero and Páramo de Sumapaz (4° S.); in equatorial Africa on Ruwenzori (0° 24′ N.), Mount Kenya (0° 12′ S.), and Mount Kilimanjaro (3° 0·5′); and in New Guinea on Ngga Poeleo (4° 05′ S.) and Mount Wilhelmina (4° 15′ S.). It is most elevated in the horse latitudes or arid, sub-tropical belt of high pressure where the precipitation is least and the temperature is highest (see below). The fall along the meridians, which is also true of Siberia,[242] is not steady but drops by jumps (fig. 116) which are related to the zone of maximum precipitation (see p. 654).

In accord with the second law, the line rises eastwards in middle and higher latitudes and westwards, but less steeply in lower latitudes[243] because the precipitation, cloudiness and humidity diminish and the summer temperatures rise into the continental interiors; the world's highest snowline lies in each of the zones, viz. at 6500 m in the Andes (25° S. Lat.) and at 6000 m on Kuen-lun (*c.* 34° N.) and over 6500 m in the Karakoram Mountains in central Asia.[244]

The easterly ascent is seen in Scandinavia, as Wahlenberg first observed (1813), in the Balkans,[245] and in the Alps whose glaciers shrink into the cirques in the east and to mere snow beds in the extreme east. It is seen too as we pass from Europe to central Asia,[246] e.g. Alps 2600–3250 m, Caucasus 2700–3900 m and Peter the Great Mountains, Transalai 4100–5000 m, and in individual mountains, e.g. the Pyrenees[247] (west, 2500 m; east, 3000 m),

Caucasus[248] (the difference is *c.* 1000 m), Kurdistan,[249] Russian Altai,[250], Kuen-lun and Karakoram,[251] Pamirs[252] and West Tienshan.[253] Alaska[254] illustrates it, as does North America generally so that the only glaciers in the east of the continent are those in the Torngat Mountains, and the Brooks Range, Rocky Mountain Front Range and Selkirk Range of the west have only a poor development of ice. The intensity of glacierisation decreases progressively on the Eurasian shelf-island groups.

The snowline, in accord with the third law, is arched over mountain ranges or plateaux. This was demonstrated for the treeline[255] of the Alps, Tatra and Scandinavia (it was first shown by K. Kasthofen in 1822), and is readily seen in the Alps[256] (see fig. 7, p. 14)—in the Swiss Alps it is computed at 600–800 m and in the eastern Alps is well shown in a map by R. v. Klebelsberg.[257] A like doming characterises Iceland[258] (coast, 600 m; interior, 700 m), Spitsbergen,[259] Caucasus,[260] Himalayas,[261] Greenland (see p. 17) and North American Cordilleras, and, as exemplified by the Caucasus, is most pronounced in the widest and highest parts. The Scandinavian snowline rises in 62° N. Lat. from 1200 to 2200 m and in 68° N. Lat. from 1000 to 1800 m[262]: the highest point is Jotunheim, the culmination of the *Massenerhebung*.

The cause of the doming is to be sought in the rise of the daily noon isotherms into the mountains[263] and the decreased precipitation and increased insolation in the same direction,[264] as illustrated in the Alps and Scandinavia. Isotherms, treelines and snowlines are higher in extensive mountainous areas and on broad plateaux than on isolated summits which behave differently regarding insolation, radiation, moisture, cloudiness and precipitation. Broad masses orographically favour snow but climatically by their very existence elevate the snowline. In some cases, the orographic factor predominates, as in ice-covered Greenland, in other cases the climatic, as in the snow-free Nevadas.

A corollary of this *Massenerhebung*, as Richter[265] noted, is that we may not calculate the snowline's height above a given mountain which is now snow-free, since the imagined greater mass would necessarily thrust the snowline upwards, nor can we, as has been done,[266] use for this purpose the interval between treeline and snowline.

As Bouguer[267] recognised, mountains if sufficiently high have besides their inferior snowline a superior one crowned by snowless summits[268]; for the precipitation diminishes upwards from its maximum (see p. 654), and absolute humidity diminishes and evaporation increases upwards too. The higher line is probably fairly low over arctic Asia and the Canadian Archipelago but, because of the condensation from clouds, may not actually exist anywhere, unless it encircles the higher nunataks in the interior of the Antarctic at about 3000 m,[269] including those of South Victoria Land[270] where glaciers are decaying from the head (see p. 419). It is apparently lacking in the heart of Greenland,[271] though the evaporation here amounts to 5·5 cm/annum.[272] It may occur in Cerro di Boneto, Bolivia,[273] and probably did so in the centre of the ice-sheets and below the Caucasian summits during the Pleistocene.[274]

Avalanches. Snow is prevented from accumulating indefinitely above the snowline by wind and evaporation (which may exceed 20% of the total[275]), by the slow draining away of glaciers, and by the catastrophic form of

ice-movement or avalanches which hurtle down from the walls (*ad vallem*). These, by removing up to one-tenth or one-quarter,[276] largely contribute to build up the névé and nourish glaciers: some glaciers are entirely so replenished.[277]

Avalanches have been exhaustively studied,[278] including their occurrence, origin and classification, the damage they do, and their forms, paths and kind of movement. The Swiss *Schnee- und Lawinenforschungskommission*, established in 1932, has made important contributions.[279]

Individual avalanches range from 200 to 20,000 cu. m, exceptionally to 2 or 3 million or even 54 million cu. m.[280] They are a normal feature in the Alps[281] (in Tyrol and Vorarlberg their annual number is 2000–3000, in the Swiss Alps, 17,500) where deforestation tends to add to their number, and in other snowy regions,[282] e.g. Norway, Pyrenees, Carpathians, Caucasus, Alaska, parts of the United States and New Zealand Alps. They play an important role in the Himalayas[283] but are rare in Greenland on account of the dry winds[284] and on mountains with gentler slopes like the German Mittelgebirge, though they are not unknown here[285]—they have been described from the Riesengebirge, Thuringia Forest and Black Forest as well as from the French Vosges. They occur on slopes as low as 22° or 14° and often move through vertical distances of 1000 m.[286]

Avalanches are usually divided into two classes, the *Staublawinen* and *Grundlawinen* (*avalanches volantes* and *avalanches terrières* of the Pyrenees[287]), terms first introduced into glacial literature from popular speech by J. Simler in the 17th century[288] and used by J. G. Altmann and J. J. Scheuchzer in the 18th century[289]; Scheuchzer also named them *Windlauwinen* and *Schloss-* and *Schlaglauwinen*. They have also been called "dry snow avalanches" and "wet snow avalanches".[290]

Newly fallen snow on high mountain sides, especially if these face east or north and are sheltered from the wind, is fine, dry and powdery[291] (Hugi's *Hochschnee*; Agassiz's *neige poudreuse*) and accumulates to a depth which depends upon the slope, the nature of the underlying surface (especially the permeability), the character of any vegetation, the composition of the snow, and the weather conditions before and after its deposition[292]—the limits of inclination may vary between 11° and 60°. In the early winter, the ground and trees may provide sufficient anchorage though long grass and flattened brush and a crust or hard-frozen under-structure may facilitate movement. A disturbance may cause a flight of the surface layers so that the "mountains smoke" or may set the underlying, heavier masses in motion to chisel countless characteristic furrows in the mantle of snow. Such "dust avalanches" (Ger. *Staublawinen*; Fr. *avalanches de poussière*; Icel. *Snöfloder*) fall at all times of the year above the snowline but below it principally during the day (when the snow-flakes lose cohesion) and the coldest seasons (*Avalanga fredda*: *Lavine di Freddo* or "cold" avalanches in the Tessin), particularly in spells of calm weather following heavy snowfalls. The slope and thickness of the snow are determinant[293] though the immediate cause may be a change of temperature, fallwinds (which begin at sunrise as exchange winds between sun and shadow slopes),[294] an acoustic vibration occasioned by a bird, shout, whistle or bell (as Scheuchzer noticed), or a disturbance of the equilibrium by a falling cornice, sérac, rock or tree, by the crossing of man or animals, or by thunder, shooting, quarrying, trains or earthquakes. The masses crash with the sound of thunder and the blast of a tornado[295]—these are the *Wind-*

lawinen of the Austrian Alps[296]—since they push the air in front and suck it in behind them: the winds may overtake the avalanche and travelling far ahead may even cause additional avalanches on the opposite slope. Air is also pressed out from the underlying snow and from the avalanche itself when this is arrested. Avalanches may attain velocities of 200 or more miles (320 km) per hour,[297] depending upon the type and volume of the snow and the smoothness and gradient of the subnival surface. Friction reduces the speed of the lateral and bottom layers.

Ground avalanches (Ger. *Grundlawinen*; Fr. *avalanches de fond*) differ from this type in several ways though numberless transitions link the two,[298] just as snow avalanches grade into wind- or snow-slabs[299] (Ger. *Windbrett, Schneebrett*; Fr. *planche*), those snow deposits of winter which lie on windward or lee slopes and have been packed by the riddling agency of moist wind and by the condensation of water-vapour. These are prone to fracture at the steepest point on a curve and slide along glide planes under the compulsion of their own stresses or if slightly overloaded by new snow, disturbed by living creatures, or subject to temperature oscillations.

Ground avalanches carry with them vast quantities of soil and rock (see p. 579), whence their name—the odours sometimes noticed in the neighbourhood of avalanches may be caused by the frictional heating of rock matter containing traces of pyrites or vegetable matter.[300] They may take place in winter but are mainly confined to the late spring or early summer (*Avalanga calda* in the Tessin) and generally to the warmest part of the day (they become progressively later as the slopes face east, south or west[301]) when the snow thaws quickly, as during the foehn. The melt-waters run off below the snows, undermine them and finally remove their support. At first, and especially if the path is smooth, the heavy wet snow (*Nassschneelawinen*) moves forward by gliding along planes which may be associated with wet layers in the snow, but later and when the path is uneven, by rolling[302]: these are Ratzel's *Rutsch-* and *Roll-Lawinen*.[303] Often the front rolls forward and the rear glides. The glissading masses are invariably associated with big gullies (Fr. *chemins*; *couloirs d'avalanches*; Ger. *Lawinenträge*) which control their course and speed. They pursue definite and regular paths (not to be confused with those thawing snows produce[304]) which frequently sweep through forests. Geological structure influences them[305] so that they are usually dangerous on limestones and schists; porous rocks delay their action.

Great cones of snow, frozen hard by regelation and made cavernous by melting, accumulate at the foot of declivities (pl. Iʙ, p. 32).[306] They assume many forms, depending upon the mode of fall, surface relief and nature of the snow. *Randklüfte*, somewhat analogous to those behind cirque-glaciers (see p. 45) and formed by movement and melting, sometimes occur at their upper surface and shear-planes may also be seen.[307] Cones of several years may combine in one locality and take years to melt.

Ground avalanches are disastrous to forests, agricultural land, roads, railways, bridges and buildings, and are menacing to life: deaths from Alpine avalanches,[308] which were over 100 in 1598 in the eastern Swiss Alps and over 300 in 1720 in the whole of the Swiss Alpine region, are roughly estimated at 10,000 per century, though during World War 1914–18 on the Austro-Italian front 10,000 soldiers lost their lives in snowslides in December 1916, in a single 24-hour period, and the total casualities from avalanches on that front were 80,000. Hence protective works and defences,[309] such as terraces,

contour trenches, rows of heavy walls, mounds, pillars, roofs, tunnels, built up galleries, and afforestation have been widely undertaken: they impede the progress of the avalanche, weaken its force, or deflect it into less harmful channels. In other cases, villages, roads, railways, power and telephone lines, etc. which are in exposed places are removed to safe positions. Where avalanche danger is cumulative, very satisfactory results have been accomplished by explosions (hand placed or pre-planted) and by shooting down avalanches in the early stages. In a winter sports area the skiers themselves prevent the build-up of dangerous avalanche conditions by constant use of the slope.

Ice-avalanches differ from the previous types since they fall from the ends of hanging glaciers, especially the steeper ones, at all seasons. The Biez, Biétroz and Glacier du Tour have furnished good examples,[310] and others have been reported from the Caucasus and other areas in recent years.[311] They may be initated by an advance, by a rise of temperature, or by waters accumulating in a glacier.[312] They range up to 4½ million or even 35–60 million cu. m,[313] and are indeed often so big that they possess ice-caves with emergent streams. The fallen masses may melt each summer or may swell the volume of a valley glacier that may be present below. If regular and persistent in their occurrence, they may also generate remanié glaciers (see p. 91) or the type which in the Andes has been called "avalanche glaciers".[314]

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Chapter II

FORM AND STRUCTURE OF GLACIERS

Névé basin. Snow which sometimes covers entire summits, as in A. Hamberg's "convex snowfields", commonly nestles in hollows, such as niches and joint cracks or big firn or névé basins. These serve as important receptacles for the snows which fall on their surface or sweep down in clouds or avalanches from the encircling heights. If such basins are generally wanting, as in Lapland and on plateaux in north Greenland, there is little firn despite the high altitude. Growth takes place also by direct fall of rain, hail, all kinds of snow, mist and hoar-frost, the latter especially in countries with a maritime climate.[1] Its annual amount has been ascertained in relatively few places,[2] as on the Hintereisferner and Mont Blanc. Observations on Mount Rainier[3] suggest that the basins receive more than the adjacent summits.

The basins are usually simple, as on the Rhône, Aar and Glarnish glaciers in the Alps, or compound as on the Aletsch and Gorner glaciers, Mer de Glace, Pasterze and Hintereisferner—the Tyrolese name *ferner* for glacier derives from the word firn. Two or more glaciers occasionally flow out of one basin as in the Hochjochferner, Oetztal, while neighbouring névés sometimes spread across the cols on to plateaux (the *champs de neige* of Agassiz).

The névé, which is free from moraines, has generally a feeble and uniform slope, modified sometimes by waves due to flow.[4] As revealed in crevasses and early recognised,[5] it is thick and more or less horizontally stratified, opaque and white bands alternating with darker ones. The white bands represent the winter's snowfall lessened by the summer's heat, while the brown or dark ones[6] enclose dust, rarely stones or rock-debris which frost loosened from the cliffs. Leaves, cones, seeds and other organic remains may be carried on to the glacier and concentrated during the summer melting. Material may also arrive from great distances. The layers are regular, unless they have been fed by avalanches or rendered discordant by uneven melting or by movements within the firn.[7] The annual layer, which forms a broad wedge, thickest towards the bergschrund and thinnest at the firnline, may be 0·5–3·0 m thick.[8]

Banding also occurs in the firn because of the retentive capacity of snow since percolating waters tend to collect in the denser layers where they freeze.[9] Melting and freezing may also cause the formation of ice-lenses and ice-glands, the former horizontal, the latter vertical, and cylindrical columns[10] (also called "pipes", "firn-pipes") which become more numerous in depth.

Few modern figures[11] are available of the firn's yearly increase, a prime factor in a glacier's economy, though various methods of measuring it have been described,[12] including gauges specially constructed and provided with shields. An efficient instrument is Mougin's snow gauge[13] though none can measure the snowfall accurately because of drift. Notable exceptions among modern figures are those from the Rhône Glacier, where a commission has fully investigated the ablation, precipitation and névé increase since 1874,[14] and certain glaciers in Iceland and Spitsbergen which H. W:son Åhlmann[15]

has examined by digging pits. On the Vatnajokull, the increase of alimentation with altitude averages 26–28 cm per 100 m, the maximum of total precipitation exceeding 600 cm. The western half of the Greenland ice-sheet above 1200 m has an annual precipitation of *c.* 350 mm,[16] the ablation decreasing from *c.* 10% at the margin to nothing at the centre.

The annual accumulation in the firn region can be simply and rapidly calculated from a knowledge of the shape, size, velocity and depth of a glacier (see p. 38). By this method, it was determined that the Rakhiot Glacier received 6–8 m of rainfall and the Fedchenko Glacier 1 m.[17]

The stratification is much less distinct in the interior of the ice-sheets of Greenland and Antarctica because, since dust and melting are generally absent, "winter" grains and "summer" grains are difficult to distinguish.

Firnification. Hottinger, De Luc, de Saussure and other early naturalists[18] described and differentiated the névé from snow and ice. Previously there had been no such distinction; the frozen mass was called *Eisberg* or *Eisgebirge*[19] long after the word *Gletscher* was first introduced in 1507. A. Heim[20] likened ice to an oolitic rock. The grains are small and spheroidal, loosely attached, and separated by air spaces. They become bigger netherwards (in central Greenland they are 0·1 mm at the surface and 1 mm at a depth of 5–6 m[21]), being fairly large at the base of the firn, as is seen where a tributary firn is hurled upon a glacier or a valley floor. Melting around the edges, shortening and eliminating of the branches, and freezing upon the modified nuclei transform the feathery skeleton crystals of the original snow into compact granular névé. This has less air, is more dense, is darker and bluer than *Hochschnee* and has a grain diameter of 0·1–0·3 mm.

Firnification takes place in several ways[22]; by the aid of hoar-frost or rime, as in Scandinavia, the coastal ranges of Alaska and Patagonia—this is of no moment in the Antarctic[23] but is important in central Greenland[24] and was probably significant on the oceanic side of the Pleistocene ice-sheets[25]; by mists, chilled by the ice, as in South Georgia, South Shetland Islands and Spitsbergen; by intense solar radiation and insolation, as in the higher mountains of the tropics and in the Antarctic; by wind pressure; and by sublimation,[26] i.e. by the evaporation of the solid ice into water-vapour, chiefly from the edges of snow-flakes, and its recondensation without passing through the liquid state, as on Jotunheim and in Spitsbergen, Greenland and Antarctica. Here the firn retains a low temperature to a considerable depth even in summer and melts very slightly, if at all, at the surface so that the ice remains bubbly—annual stratification is unknown in the heart of the Antarctic and the normal grain size at *Eismitte* in Greenland was 1 mm. The most important factor of all is melting and refreezing of percolating thaw water.[27] The rate of downward percolation in a firn depends on the rate of water-supply, the crystal size and density of the firn, and the thickness and freezing of ice-bands. The latter may hold up the vertical flow of water and cause it to move with a marked horizontal component[28] though escape is found through flaws or thinner parts of the bands. In a different climatic environment compaction-settling may be more important[29]: it acts at all seasons and at all depths. Where melt-water is lacking it may be the sole mechanism[30].

Laboratory experiments demonstrate that the growth increases with temperature and is most rapid at or near freezing point.[31] Thus the snow changes quickly into firn on tropical summits, e.g. Kilimanjaro,[32] and may

be so transformed in a single day[33] in the warmer air of the Alps and Norway. By contrast, the particles in the very cold Ross Barrier are still angular and incompletely crystalline after two years.[34] Incoherent snow occurs without melt horizons down to 4 m in central Greenland[35] where the temperature very rarely reaches melting point and there is neither condensation nor evaporation. Measurements show that the density, which is much less than in Spitsbergen, rises fairly rapidly to a depth of 7 m where the uncompacted layers cease and the rate of increase becomes less.[36] On the Antarctic plateau south of 88° 25′ Lat. soft snow has been penetrated to almost 2 m.[37] But below this, as seismic methods prove (see p. 39), the density increases as does the elastic property of the firn,[38] though this may be 60–70 m deep, as in Greenland.[39]

Contact with glacier-, lake- or sea-ice largely facilitates granular growth.[40] The change progresses upwards from the junction. The ice grows at the expense of the overlying snow, though the determinants in the case of sea-ice are the relatively high temperature of the upper surface and the quantity of salt caught up in the crystals.

Glacier grains and glacification. Firn ice (*glace de névé* [41]) consists of rounded grains, rarely larger than 2 mm, with interstitial finer fillings and with much imprisoned air which is irregularly and finely distributed.[42] It derives from the firn when percolating waters freeze and passes into glacier ice (the change may take place without its intervention[43]) from which it differs since it possesses an ice-cement, lacks the system of communicating capillaries, and is only about two-thirds as dense.[44]

The transition from snow to névé and glacier-ice, i.e. from the "tesselate" névé (equidimensional and unsutured) to the sutured and inequigranular ice,[45] is accompanied by a rise in density[46]: 1 cu. m of snow weighs 85 kg, of firn-ice 500–600 kg, and of glacier-ice 900–960 kg. It has been followed[47] for example on Mont Blanc, where glacier-ice with a density of 0·86 and a grain diameter of 2 mm was found at a depth of 15 m, on the Tête Rousse, Mer de Glace, the Claridenfirn and in the heart of Greenland, though here the depth of the change can be calculated by extrapolating the specific gravity curves (see below). In west Greenland, the density was 0·53 at a depth of 5 m.[48] On the Jungfrau,[49] the physical properties at successive stages were as set out in the table below.

S. Gr.	*Grain Size*	*Depth*	
0·291	0·1–0·2 mm	0·16 m	Settled snow
0·437	0·3–0·5 mm	1·10 m	Early firn
0·463	0·6–1·3 mm	2·70 m	Firn
0·690	1·0–2·0 mm	23·25 m	Denser firn

On the Sonnblick,[50] the density at 12·5 cm was 0·236 and at 187·5 cm 0·795.

The critical density at which the intercommunicating air ceases to exist and the firn becomes ice varies between 0·82 and 0·84.[51]

The increasing compactness results from the "metamorphism" or "diagenesis" of the snow,[52] i.e. the gradual growth, readjustment and closer packing of the crystals and the expulsion of the entangled air[53]; for while freshly fallen snow has up to 97% of air (the specific gravity is about 0·1 but may be as little as 0·01–0·05[54]), the amount being less if the snow is wind-driven,[55] the percentage is only 55 in Tyrolean névé,[56] 6 in the white ice-layers,[57] and

2·5 in the windy Antarctic[58] where in the interior the snow is in consequence usually hard and closely packed. The increasing compactness is borne out too by the rise in the pollen density of the ice throughout its course.[59]

This process of compacting is intimately connected with movement, i.e. with rotation and relative slip. When discontinuous, it causes a sudden collapse over considerable areas,[60] as during sledging and skiing. The imprisoned air escapes quietly or with a slight whistling or hissing, or with a dull thunder-like noise or crash[61] (Ger. *Firnstoss*), as on Ross Barrier and in Greenland where the noise and "snow tremor" startle eskimo dogs and ponies. The depth of the firn, which has a stratification determinable by density changes, has been ascertained in Greenland by seismic methods (see p. 39) to be 300–350 m 62 km from the ice-margin.[62]

The air is held in the ice as bubbles, more or less drawn out in the direction of flow and under pressure ("parallel" or "fluidal" texture of A. Hamberg[63]). This amounts to 10 atmospheres 7·5 m down in the cold ice of Greenland[64] and causes icebergs to burst asunder and crumble suddenly into innumerable fragments.[65] It is released[66] on warm days and from the walls at the opening of crevasses and when blue bands are forming, as well as by melt-water, by liquefaction and subsequent regelation, and by the expansion which accompanies freezing. It would seem that there is no mechanism for eliminating the air bubbles from glacier-ice, except in narrow bands which become clear of air by soaking in water at the névé stage.[67]

Glacier-grains are a vital factor in ice-motion which is associated with the mechanism of their growth (see ch. V). Their investigation constitutes the most material advance in glacier studies since the labours of Agassiz, Forbes and Tyndall in the fourth and fifth decades of the last century. Much, however, still requires to be done on the changes in the size and orientation of grains, in the pressure, distribution and elimination of air-bubbles, and in the physical chemistry of the liquid phase associated with the bubbles and the granular margins.

Glacier-ice, including that in reconstructed[68] and ice-foot glaciers,[69] is a confused crystalline aggregate as W. G. Ploucquet and Hugi were the first to note.[70] The grains are curved polyhedra, tightly interlocked,[71] and are of irregular size and shape and lack any crystallographic order[72] (fig. 9). The grain structure may be studied by paper and pencil rubbing on melting ice,[73] by using a clay and absorbent paper,[74] or by a formvar replica technique.[75] Each grain is a single crystal as was shown by early observations in polarised light,[76] by Tyndall's melt-figures (see p. 58), by Forel's stripes (see p. 58), and by experiments,[77] though the surfaces are not crystal planes, as is the case in river- and lake-ice. The Hugi effect (see p. 57), infra-red absorption and the phenomenon of regelation suggest that the molecules at the granular contacts are in an amorphous state.

The grains grow bigger towards a glacier's edge, snout and base: those at the margin may be larger because they have been more exposed to daily fluctuations of temperature caused by radiation from the walls, and because they have travelled for a longer time and are therefore older.[78] Hence the grains are tiny in short or steep glaciers,[79] as "large as melons" in the long glaciers of Greenland,[80] largest (and often elongated) in dead ice[81]—laboratory experiments show that crystals grow rapidly after the release of stresses,[82] though the large size in dead ice is also attributed to melting and refreezing[83]—and differ in size in compound glaciers.[84] The mean annual increase in diameter

is 1·4% in the Alps.[85] The grain-diameters[86] here range up to 7·6 cm, in Spitsbergen to 10·2 cm, in Grinnell Land up to 3·8 cm, in Greenland to the size of small peas, hazel nuts, walnuts or pears, in the Canadian Rocky Mountains to 7·6 cm, in Franz Josef Land to 1·5 cm and at Cape Adare, Antarctica, to 0·75–1·5 cm. The mean in the Antarctic has been judged to be 0·2 cm.[87] Recent observations suggest that under certain conditions they are surprisingly small, depending possibly upon the inclination of the glacier-bed[88]: rapid movement keeps the grains small.[89]

While the transformation from snow crystals, the germs of glacier-grains, to glacier-ice styled "glacification" by Agassiz[90] or "firnification" by Seligman,[91] may be taken for granted (T. Vidalin noted it as early as 1695), exactly how it is accomplished has rarely been observed[92] or experimented upon.[93] Consequently it is only imperfectly understood: the driving down of the stratification and streamlines in the firn and their rising about the firnline (see p. 119) explain why the grains fashioned in the depth of the firn emerge as completed coarse grains without an observable transition.[94]

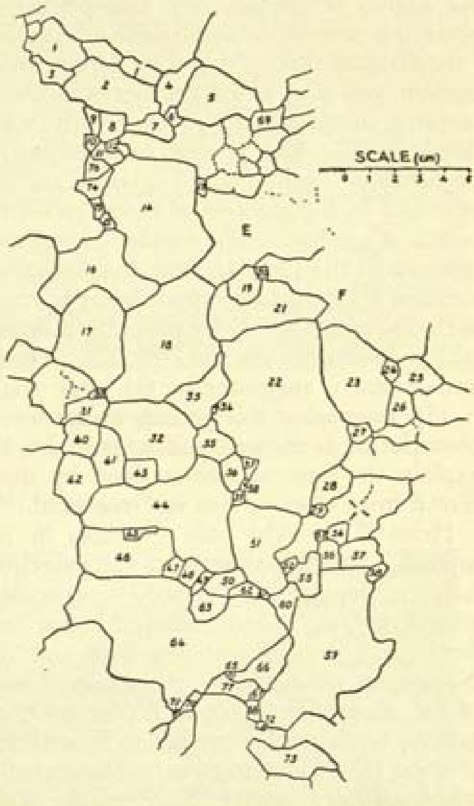


FIG. 9.—Glacier-grains from the inland ice at Eqe, West Greenland. M. Boyé and A. Cailleux, *J. Gl.* 2, 1954, p. 325, fig. 2.

Snow and névé change into ice with remarkable uniformity at a density of between 0·82 and 0·84 by losing the distinctive cement and by developing a capillary net. The grains crowd together and become homogeneous and transparent. That this may take place at a shallow depth is proved, for example, in tunnels through the névés and glaciers in the Tyrolese Alps[95] and on Mont Blanc where transparent granules were encountered only 4 m down.[96] In central Norway the depth was 3–5 m.[97]

Ordinary snow is converted into firn snow by settling and packing and by the transfer of water from small crystals and pointed parts of crystal flakes to the larger particles. The process, which involves a reduction of the total volume, is one of change from highly branched crystals to crystals much more nearly approaching the spherical shape. It is facilitated by wind[98] which apart from its fluctuating pressures causes a constant "breathing" of air in and out of the snow to a depth possibly of many metres, as in Ross Barrier, and, especially if the air is moist, a transference of water-vapour.

A large percentage of the firn crystals near the surface of the névé have their horizontal axes at right angles to the surface of the glacier, i.e. parallel

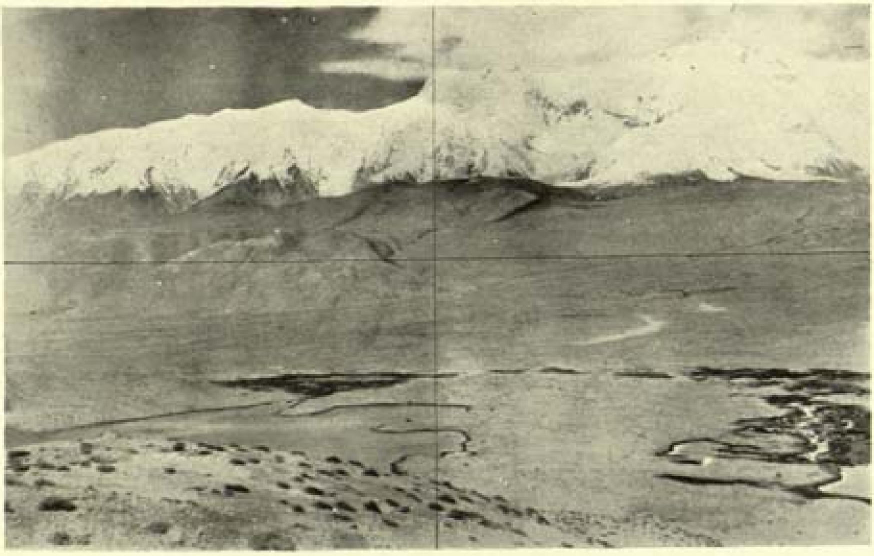
to the direction of the temperature gradient.[99] This initial orientation is destroyed in depth, e.g. at 23 m, by the glacier's differential flow which causes a redistribution, though it is retained in the ice-bands in which dense packing prevents independent movement.

Further change is brought about by percolation and infiltration[100] (Forel's "thermal theory"). This is only true of the outermost layers (*Verdichtungszone*[101]) because percolation is only possible for short distances in cold glacier-ice[102] and is controlled by the impervious ice-bands—pollen infiltration from the higher layers does not take place.[103] It is affected too by new layers accreting around grains which themselves undergo no change[104] (Forel's "mechanical theory") and by larger grains growing at the expense[105] of the cement and of strained grains (as in the case of metals) and, as experiments confirm, of the smaller grains[106] which finally disappear—they long escaped detection.[107] The process is aided at the surface by evaporation and condensation[108] (sublimation), by rain and melt-waters, and in depth by plastic flow and by liquefaction at points of contact and by subsequent refreezing at points of tension. The smaller grains are eliminated by molecular transpiration across the films and the higher vapour pressure[109] and curvature of the surfaces,[110] the water-vapour being transformed from the smaller to the larger surfaces, and possibly by granules disintegrating into their constituent plates which constantly reunite.[111] The importance of radiation (heat and light) and of time is suggested by the large grain size in dead ice.[112]

The reversal of this process under severe strain,[113] as when ice encounters obstacles or is sheared and large grains break down into smaller ones, may explain the limiting size attained by the grains (see above) which was inferred from observations in Greenland.[114]

Heim[115] thought only granules in parallel crystallographic orientation united. Parallelism between the principal axes and the direction of pressure has been repeatedly affirmed,[116] especially for the basal layers and the snout where shearing is most marked. For instance, the grains in Greenland are flatter and parallel with the floor and sides of the glaciers[117] and in some Norwegian glaciers have their Forel's stripes (see p. 58) in the shear planes of the snouts.[118] Certain Alpine tongues show a granular orientation or a general tendency for crystals to lie with the basal planes parallel to the planes of shear.[119] Laboratory experiments confirm the arrangement of the crystals when subject to stress.[120] Even the grains in the firn may be orientated in some degree by downward growth.[121] Adjacent grains showing a parallelism of their Forel's stripes have been thought to have been broken down from one grain.[122] It is, however, contended that there is little or no preferred orientation[123] and, if present, it is too rare to explain granular growth. But, as G. Tammann has shown, a first requirement of such growth is that the separating layers of salt solution become so thin that the grains come into contact with one another and that the surfaces in contact are crystallographically equivalent. Growth takes place especially in the deeper parts, glacier movement being an essential condition: the greater velocity, the quicker the growth. Crystals whose orientation allows them to yield to shearing forces by gliding along basal planes are subject to smaller stress than other crystals and having less free energy grow by transferring molecules across their boundaries.

The importance of pressure in granular growth was early recognised.[124] It is proved experimentally,[125] as well as by the large grains which occur



A. Snowline in mountains of China [Royal Geographical Society]

B. Avalanche cone, Paznaun, Austria [W. Flaig collection]

A. Corrie glaciers with bergschrunds, Hintere Schwärze, Austria [W. Flaig collection]

B. Randkluft, Hochwilde, Gürgler, Ötztal [Transmitted by H. Louis]

(*facing p.* 33)

where the pressure fluctuates,[126] by sections through avalanches,[127] and by the conversion into glacier-ice of snow which falls into crevasses.[128] "Glacification" by "dry union" well below freezing point[129] is confirmatory.

Observations[130] in Spitsbergen, west Greenland and Franz Josef Land show that higher pressures are needed in the cold Arctic to convert névé into glacier-ice. Hence many arctic glaciers, as early noticed,[131] are not thoroughly consolidated but exist as firn-ice or "undeveloped glaciers" (*glaciers imparfaits*[132]). In the Antarctic, where temperatures are constantly low and variations few, much of the surface consists of frozen firn and the grains do not grow or remain small.[133] The ice is less dense and blue than in the Alps, is porous and penetrated by countless air-bubbles,[134] and is possibly more plastic[135] (cf. p. 48). This slow growth explains too the relatively small grains of polar glaciers[136] despite their greater lengths.

Hence infiltration is important in the névé but glacification is mainly accomplished by pressure melting, large grains growing by eliminating the cement and smaller grains.

While crystallisation in temperate glaciers with large ablation is rapid and converts the firn into ice at slight depths, the boundary between the two being sharp, subpolar glaciers consist in their higher parts of hard crystalline firn, formed by slowly recrystallising the annual surplus of the accumulated solid precipitation that passes gradually downwards into ice. In the polar type of Greenland and Antarctica, this transformation into firn is too deep to be observed, the firn having a temperature that is always negative and even in summer is so low that melting does not take place and any change is primarily by sublimation. In Spitsbergen, whose mean summer temperature is about 0°C, the firn, proved to a depth of 15 m,[137] is probably only about 20–30 m thick.

Glaciers without firn. Active glaciers without firn have occasionally been noted in the Alps[138] and the Arctic[139] and much more commonly in middle latitudes in Asia[140] (Caucasus, Turkestan, Himalayas, Karakoram, Pamirs) where they may be exceptional or the rule; they are variously termed the "firn-basin"[141] (*Firnkessel*), "Turkestan",[142] "Bogdo-Ola"[143] or "Mustagh"[144] type. The Barnes Ice-cap on Baffin Island, which is *c.* 6000 sq. km in extent and appears to have an approximately balanced budget, has no firn: P. D. Baird has proposed the name "Baffin-type" for such glaciers (see p. 80) which lie below the snowline and have great residual cold, light precipitation and short cool summers and long cold winters.

Exceptional melting of the firn during warm summers explains this peculiar type in the Arctic. Elsewhere, the high, steep rock-walls produced by drastic preglacial rejuvenation cause the glaciers to be fed in the "regeneration region" (H. Ficker) by avalanches or hanging névés in winter and spring, and by ice-avalanches at all seasons. They are, therefore, concave, lying partly or completely below the snowline[145] and have been termed "avalanche glaciers".[146] The Pleistocene glaciers in the Karwendelgebirge were apparently of this kind.[147]

Glacier tongue. The division of a glacier into firn basin and glacier tongue (*glacier proprement dit* of Agassiz[148]), *Sneebrae* and *Isbrae* of the Danes, was first made by Simler in 1574[149]: W. Flaig[150] has given a full account of the nomenclature and history of the earlier *Gletscherkunde*. The two parts are readily recognisable from a distance: the firn is white, concave and

relatively broad; the tongue is blue, convex, narrow and long, though occasionally, as in the Übergossene Alm,[151] it may be much broader than long. The firn is the seat of accumulation (Rendu's "reservoir"[152]), the tongue of ablation and loss (Rendu's *glacier d'écoulement*, Reid's "dissipator"[153]).

The mean slope is usually less in the tongue than in the firn and is generally steeper the shorter the glacier; on the Aletsch it is 4°, on the Mer de Glace 5–6°, and on the Monte Rosa Glacier 10–20°. In the Alps,[154] it averages 3–6° but varies between 1 in 20 and 1 in 2: the hanging glaciers vary between 25°–30°, the larger glaciers between 5–7°. Other figures[155] are as follows: Fedchenko 2·8–3·2%, Rakhiot 10%, Chongra 11·4%, Sachen 8·3%, Bazhin 7·8%, Zemu 4% and Baltoro 2·3%. The average slope of the outlet glaciers of Antarctica is as follows[156]: Thorne 1 in 66, Beardmore 1 in 90, Koettlitz 1 in 40, Ferrar 1 in 45, Taylor 1 in 25, Mackay 1 in 25 and Drygalski 1 in 90.

The slope as a rule gets ever steeper towards the snout owing to increased ablation. Reversed slopes with small lakes and backward flowing streams occur occasionally[157] if, for example, a glacier rises over a swell or piles up against an obstacle.

Convexity. The transverse convexity, noticed by Gruner,[158] is greatest in smaller glaciers[159]; the middle of the Aletsch Glacier was in 1872 60 m higher than the sides.[160] Since radiation from the rock-walls is the main cause, large glaciers are almost flat from side to side; the sides on north–south glaciers suffer equally; and on east–west glaciers[161] in the northern hemisphere the southern flank is chiefly affected, as on the Aar and many Himalayan and Karakoram glaciers, and south of the equator the northern flank. The convexity is most pronounced in the tropics and subtropics,[162] as in the Himalayas and Pamirs. It would be greater but for the protecting avalanched snows and morainic cover, and for the slight compensatory movement to the sides.

An added factor is the increased evaporation towards the axis which raises the temperature of the ice and lessens its rigidity. This may explain why glaciers in South Victoria Land constantly tend to shift their centre of activity towards the north,[163] the south side being relatively inert.

A glacier's flanks rarley fit flush against the rock-walls; the Little Tokiehitna Glacier in Alaska[164] is a rare instance. Almost invariably, even on the outside of bends, there are radiation gullies which diminish upwards to the snowline where they cease. Kept open by radiation and running water and by unmelted avalanches which conduct the frost debris out on to the glacier,[165] the gullies are largest in the tropics and subtropics, as in the Karakoram's east–west "ablation valleys"[166] where the ice-edge, even on the convex sides of bends, may rise sheer through 50 m: radiation is high and warm waters issue from the valley sides—these gullies may also be due to ice-flow (*Bewegungsrandklüfte*[167]) as well as to ablation.

The convexity is also partly accounted for by the quicker flow of the ice along the median line (pitch and other viscous bodies furnish analogues); for the line of maximum elevation is roughly that of maximum flow[168] and the lessened flow during the present retrogression (see ch. VI) has lessened the convexity.[169] Additional factors are the rigid outer layers which may play a role similar to surface tension in a liquid,[170] and lateral moraines if these are not too big to be protective.

The diving down of the stratification and line of maximum flow in the

névé may partly explain the concavity of the firn while the rising of the stratification and line of maximum flow in the tongue may be partly responsible for the tongue's convexity.[171]

Yet glaciers may be flat or even concave[172] if they are replenished along their sides (see p.33) or their lateral moraines afford exceptional cover (this may be on one side only as on the Vernagtferner). Glaciers in dry climates with energetic evaporation[173] also have this shape, as have those which belong to the type of rapidly moving tidal glaciers which float in their central parts and rest on land on their margins.[174] In some areas, as in north-east Greenland,[175] compound glaciers may be traversed by pronounced longitudinal furrows running for miles up and down that delineate the several units.

Termination. The thicker ice and more rapid flow on the axis make the snout convex downstream, though the incidence of the sun's rays may control the actual shape.[176] Occasionally, it is a re-entrant angle, as on the Lower Grindelwald Glacier in winter,[177] or in ice-fjords[178] where the flanks rest on land or in shallow water and the centre floats and wastes rapidly by calving and, to a less extent, by melting. Certain glacier-lakes of the Pleistocene also had concave ice-faces (see p. 457).

The terminal slope varies considerably. In tide-water glaciers which have suffered little from abrasion it is usually vertical unless, as often in north Greenland, e.g. in Petermann Fjord and Sherard Osborne Fjord, the sea is constantly frozen and the ice slopes gently.[179] On land, the snout may taper or be vertical. Slopes rising gradually backwards are limited to regions where, as in most Alpine glaciers and in south Greenland, surface ablation plays a large part[180] and loss exceeds supply, the difference between the factors increasing steadily towards the snout. The steepened slope behind the latter is, however, also partly due to the building of the ice behind the thin and rigid snout (see p. 119) to enable this obstruction to be overcome, especially in valleys with low gradients.[181] Stagnant glaciers without supply have smooth convex surfaces, flat terminations and feather edges.

Vertical snouts have been frequently described from polar regions,[182] including north Greenland—the Upernavik Glacier has the highest cliff[183] (100 m)—Spitsbergen and the Antarctic, and vertical faces around nunataks.[184] The cliff may have an overhanging cornice and a notable talus,[185] made of pieces detached from the top or of debris melted out from the base.

These "Chinese walls", as Lockwood[186] styled them in Ellesmere Land, are due to quicker flow in the upper layers[187] accompanied by shearing and marginal thrusting; to retardation or more rapid melting of the lower, debris-laden ice,[188] the overhang being proportionate to the quantity of detritus; to washing and undermining of the base by melt-waters and glacier-streams[189]; and to a glacier's greater thickness which ensures that the internal temperature is below freezing point.[190]

The difference between receding faces and cliffs may be one of plasticity[191] or may be due to latitude[192]: where this is high and the sun is above the horizon for long periods, the rays are incident at a low angle and mostly fall upon the glacier's edge. But latitude is not the main factor; for tapering snouts[193] occur in Ellesmere Land and North-East Land and vertical faces[194] in the Pamirs, Karakoram Mountains, Alps and on Kilimanjaro. Occasionally, a single glacier may have the two kinds of termination.[195] High latitude works mainly in ensuring low temperatures and slight melting,

notably if surrounding mountains provide protection from the sun's rays.[196] Philipp[197] sees its effect in the proportion of direct and indirect ablation: temperate glaciers melt by direct ablation, arctic glaciers principally by insolation and indirect ablation acting by means of the dust cover. Since ablation in the Arctic increases upwards, the glacier's thickness remains more constant. In the other zone, it increases downwards to a tapering snout.

Although these various factors doubtless contribute, the main cause appears to be the glacier's state,[198] i.e. whether it is in retreat or advancing, as immediately controlled by the flow of its upper layers. The Grindelwalders early anticipated this explanation—an advancing glacier had its nose in the air, a receding one its nose in the ground.[199]

K. E. v. Baer[200] concluded in 1860 from observations in Europe and Asiatic Russia that northward flowing rivers in the northern hemisphere were deflected to the right by the earth's rotation—J. Babinet[201] had anticipated this conclusion by one year and P. A. Slovtzov[202] by ten years. While the force is not to be denied, its adequacy is less certain. Most writers who have examined the matter have accepted the law,[203] and attempts have been made to confirm it experimentally.[204] Yet others[205] doubt or repudiate it, thinking its effects are masked by more potent factors, e.g. hardness of rock, permanent winds, tectonic movements, fan building by tributaries, or human influences.

Glaciers with their sluggish flow are not meridionally deflected. The occasional instances which have been observed are due either to quicker flow in the borders which are more exposed to the sun's rays[206] or to the markedly unequal size and strength of enclosing morainic ridges.[207]

Thickness of modern glaciers. The thickness of glaciers and the configuration of their beds are very important for glaciology, geophysics and civil engineering. Few data are available of the thickness of glaciers, fewer still of névé basins. Direct observation is only possible if a glacier plunges over a rock-face or ends in a steep snout or if retreat uncovers the floor. Examination of areas recently vacated by ice-retreat shows that ice-tongues were within the last century as much as 214–244 m thick.[208] Soundings in the fjords[209] in front of retreating Alaskan glaciers showed that the Nunatak Glacier between 1906 and 1909 was 230 m thick, the Muir Glacier 9 km from its end of 1892 was 725 m thick and the Grand Pacific Glacier was 760 m thick at a point 19·5 km from its termination in 1894. Soundings in front of the Rink Glacier of Greenland gave 700–800 m.[210]

Alpine crevasses (cf. p. 47) have given the following depths[211]: Mönch and Jungfrau firn, 28 m; Rhône Glacier and Mont Blanc, 33 m; Glacier du Bois, *c.* 30 m; Tête Rousse, 41 m; Lower Grindelwald Glacier, 52 m; and Aar Glacier up to 60 m or 260 m—the average depth in Alpine or other temperate glaciers may be correlated with that of the known yield stress of ice.[212]

Moulins supply comparable figures[213]; for example, the Pasterze, 31 m; Lower Aar glacier, 39 m; Mer de Glace, 68 m; Lower Grindelwald Glacier, 105 m; and Finsteraar Glacier, 232 m. In Alpine glaciers, they are generally less than 100 m.

Unfortunately, crevasses and moulins rarely reach the base (see p. 47), except in hanging glaciers. Hence glacialists have in the past been forced back upon estimates[214] such as 200 m for the Mer de Glace, 300 m for the

Aar and Malaspina glaciers, 400–500 m for the Jostedalsbrae, and 600 m for the Tasman Glacier, New Zealand—S. Münster[215] in his *Cosmographica universalis* of 1543 had already suggested a depth of 300–400 fathoms. Alpine glaciers[216] have been thought to be 400–500 m deep and in the great basins several 100 m. These estimates may be checked by figures supplied by the recession of the alpine glaciers since 1850, by bores (see below), by seismic determinations and by calculations connected with the glacier's economy.

After Agassiz sank his shallow bores of 60 m on the Lower Aar Glacier[217] nothing was attempted in this way until the present century when they were drilled[218] in the Mont Blanc region, in Greenland, in the Malaspina Glacier and in Swedish Lapland. The tongue of the Hintereisferner[219] was pierced between 1894 and 1910 in 12 places and to a maximum depth of 224 m. From its figures Hess calculated the precipitation in the firn and the velocity and ablation, and constructed a longitudinal profile, a map of the bed and several cross-sections.[220] He likewise attempted from published data to draw the profiles of the Rhône Glacier.[221] Boring by use of bodies heated electrically or by boiling waters[222] have generally not proved successful though a depth of up to 195 m has been reached and electrothermic sounding has been found useful on French and other glaciers,[223] including the Malaspina Glacier where a depth of 1950 ft (*c.* 455 m) has been proved—the rate may be up to 25 m/hour with a diameter of bore of 60–80 mm. When the hole reaches bedrock it usually empties of water and the subglacial stream can sometimes be heard. Such waters are now being prospected for hydroelectric plants in high mountainous country.

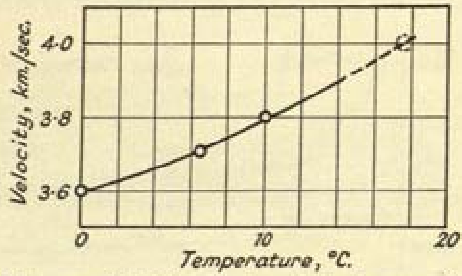


FIG. 10.—Velocity of longitudinal waves in glacier-ice according to the temperature. H. Brockamp, *N. J.* 93, 1951, p. 215, fig. 6.

Deep bores through moving and crevassed glaciers need heavy and cumbersome equipment and are costly and technically difficult.[224] The new seismic acoustic method is therefore a welcome substitute. Using a vertical seismograph with optical registering, H. Mothes[225] was able to determine the rate of propagation of the shock-waves from explosions through the Hintereisferner and, by obtaining their point of reflection from the junction of ice and rock, to calculate its depth. The velocities of propagation through ice of this and other glaciers has been found to be 3·4–3·6 km/sec for longitudinal waves and 1·6–1·69 km/sec for transverse waves and to vary with the temperature of the ice[226] (fig. 10), the reflection curve being a parabola—the velocities near the less compact firn were smaller. The computed depths obtained by Mothes equalled exactly those given by Hess's bores, after allowing for an intervening ablation of 15–23 m. The seismic method gave 792 m and *c.* 500 m for the Great Aletsch Glacier[227] (Concordiaplatz), *c.* 321 m for the Pasterze,[228] 237 m for the Rhône Glacier,[229] 444 m for the Lower Aar Glacier[230] and 95 m and 200 m for the Stor Glacier of Swedish Lapland.[231] A thousand measurements on the Gorner Glacier, which yielded a maximum depth of 450 m, gave for the first time the complete map of a glacier-bed.[232] Figures have been obtained for other Alpine glaciers and for the Vatnajökull (740 m) and Malaspina Glacier.[233] The seismic depth of the ice in Baffin

Land is 1300 ft (P. D. Baird, 1953) and of the Saskatchewan Glacier, Alberta, 1450 ft (M. F. Meier, 1954).

The weight and complexity of this equipment, as well as the need to use explosives, have led to a preference for electrodynamic methods, depending upon the electrical resistance of ice and snow and the ground beneath[234]—these have been applied for example, to the Pasterze[235] and the Crillon Glacier, Alaska[236] (270 m).

If ice behaves like a highly viscous liquid (an assumption which, though erroneous, gives approximate results) the depth may be calculated provided gliding is negligible and the velocity, viscosity and the inclination of a glacier's surface are known.[237] Lagally's development of Somigliana's theory serves as a working hypothesis,[238] though seemingly only in stationary glaciers[239] (see p. 121)—it is not applicable to glaciers which have a block movement (see p. 118).

The verification of Lagally's and Somigliana's theories by the subsequent seismic soundings of Mothes and Brockamp (see above) established the theory of continuous, stationary, viscous flow, provided that discontinuities are

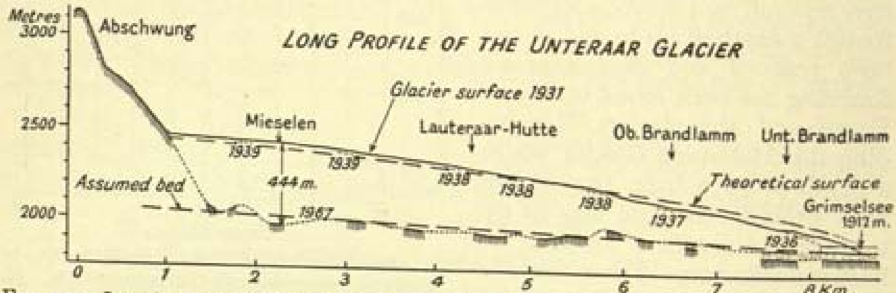


FIG. 11.—Longitudinal profile of the Lower Aar Glacier measured by seismic sounding (P. L. Mercanton and A. Renaud) and the curve calculated by J. N. Nye. Vertical scale twice the horizontal. J. N. Nye, *J. Gl.* 2, 1952, p. 103, fig. 1.

absent. Determination of the glacier economy, speed measurements with modern theodolites and by photogrammetry and pollen analyses have confirmed the main results on slowly moving glaciers. The thickness calculated by assuming the valley of the Lower Aar Glacier to be a cylinder of parabolic cross-section and the shear stress on the bed to be constant and equal to 0·77 bars agreed closely with that obtained by seismic soundings for this glacier[240] (fig. 11).

By this method the Fedchenko Glacier gave 550 m,[241] the Rakhiot Glacier 250 m,[242] the Goose Glacier, Spitsbergen 230 m,[243] the Pasterze 280 m and the Obersulzbachferner 140–45 m,[244] the Styggedalsbrae 105 m[245] and the Lewis Glacier on Mount Kenya 60 m.[246] The thickness has also been determined by applying Navier-Stokes' equations[247] for viscous streaming. Krueger[248] applied to parts of Greenland I. Högbom's theory of abrasion and plucking as given by the fineness or coarseness of the material delivered subglacially by the glacier.

Thickness of modern ice-sheets. The difficulties in finding the thickness are enhanced where ice-sheets almost completely bury a land of unknown shape. Greenland is an excellent example. It has been held to be a congeries of ice-girt islands,[249] e.g. by C. L. Giesecke and W. Scoresby, an ice-filled bowl within a ring of mountains[250]—this is in accord with recent calculations

based on the magnitude of the inward shear force exerted by the rock floor, which suggests that the floor is 1200 m below sea-level—or a dissected plateau like Norway.[251] Its control of the ice over a considerable distance from the margin is evinced by the fjords' inward extension as depressions in the ice-surface,[252] sometimes with precipitous and fissured heads, e.g. in the Ryder, Petermann and Ostenfeld glaciers and in the Jakobshavn Glacier where the effect is discernible 145 km from the western edge. It is also seen in the gigantic steps,[253] sometimes 80 km wide—3 km, 9 km, 17 km and 27 km seem to be usual[254]—which seismic methods prove are related to the underlying relief, and which pass in the interior into flat and uncrevassed waves or swells—they have been credited to the snow-broom.[255] The terraces usually slope outwards but fall inwards in the north-east and occasionally elsewhere[256] and give rise at lower levels to elongated lakes.

Estimates of the central thickness vary very much according to our view of the subglacial relief: figures range from 300 m to 2133 m.[257] W. Meinardus,[258] using a more scientific method (see below), found the rock-floor averaged 400 m in altitude and the surface 1800 m, the thickness being 1400 m. The

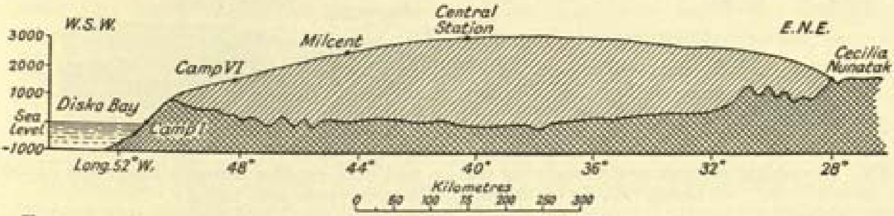


FIG. 12.—Seismic profile across Greenland. J. J. Holtzscherer, *G. J.* 120, 1954, fig. 6.

depth is much more in the valleys (1980 m and 1000 m have been mentioned[259]). The mean height of the ice-surface was found by F. Loewe to be 2100 m, by A. Bauer (1955) to be 2135 m and by G. C. Simpson 852 m. Thicknesses of the ice[260] of 1·95 km have been obtained by gravity determinations and of 2·5 km by other methods.

Dynamite explosions in Greenland,[261] with charges of up to 28 lb at suitable distances of one or two kilometres, i.e. about twice the thickness of the ice where total reflection from the rock occurs, have given depths of 1220 m (at 40 km from the ice-edge and a height of 1500 m), of 1850 m at a distance of 120 km, and of 1900 m ±130 m at a distance of 400 km at *Eismitte* (70° 53′·8 N. Lat., 40° 42′·1 W. Long.) at the geometric centre of the ice-sheet (perhaps 160 km west of the ice-divide), where the surface is 3030 m ±20 m high[262], and of *c.* 3000 m at the Central Station of the *Expédition Polaire Française* in the area corresponding to the central dome (see p. 75). These depths are somewhat uncertain: apart from technical errors, they ignore the layers of different density in the ice and assume that the reflection and surface points are in the same plane and that surface and reflecting planes are parallel. The thickness of the firn was found to be 50 m at a distance from the edge of 42 km, 105 m at 62 km and 145 m at 120 km and at *Eismitte c.* 300–350 m.[263]

That the Antarctic ice is thin on the plateau is seemingly indicated by crevasses associated with pressure-ridges[264] and by variations in the magnetic declination due to local attractions by thinly buried land masses.[265] Nevertheless, thicknesses up to 3050 m (10,000 ft) have been mentioned.[266] Meinardus,[267] who contrasted the concave hyposographic curves of the

continents with the convex curves of Greenland and the Antarctic and ascribed the difference to the ice-sheets, calculated from meteorological conditions the Antarctic's average height to be 2,200 m, a figure somewhat higher than that Barlow[268] obtained. Assuming the concealed land had the average height of the neighbouring continents of Africa, South America and Australia, viz. 650 m, he obtained a mean thickness of *c.* 1600 m for the ice. Seismic methods have proved up to 640 m back of the Alexandra Mountains[269] and later a depth of up to 2400 m[270] as in Neu-Schwabenland. The thickness of the ice in Dronning Maud Land at an altitude of 1590 m is 2175 m; ice-fjords, some 600 m deep, penetrate more than 300 km inland (V. Schytt, 1954).

A. E. Nordenskiöld estimated the thickness in North-East Land at 60–90 m (cf. p. 80). The ice-domes of Greenland (see p. 74) may be related to a thinly buried relief[271] or to exceptionally heavy precipitation[272] (see p. 664).

Thickness of Pleistocene ice-sheets. It might be thought that if the difficulties attending the determination of the thickness of living ice-sheets

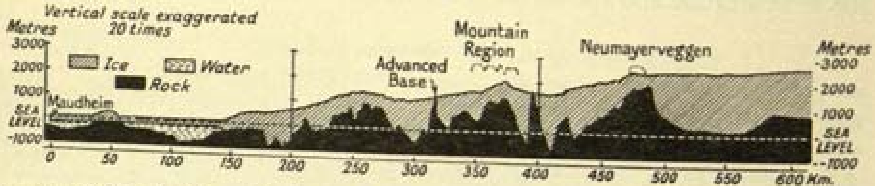


FIG. 13.—Main seismic profile showing the ice and rock profiles through Dronning Maud Land. G. de Q. Robin, *J. Gl.* 2, 1953, p. 208, fig. 4.

are so considerable, the Pleistocene ice-sheets must present well-nigh insuperable difficulties. That this is not so we owe to certain fortunate avenues of approach.

Existing ice-sheets afford little help: their depths are uncertain and direct comparison is inadmissible. They cannot be taken as norms[273]; their nourishment and wastage are insignificant compared with those of the Ice Age; they are controlled by smaller radiation, lower temperatures and lighter precipitation; and their gradients are lower. Nevertheless, since the height as proved by modern Greenland and Antarctica increases but slightly with greatly lengthened radius (see p. 44), we may conclude that the Pleistocene ice-sheets had approximately the same height.

The problem fortunately can be attacked by supplementary means. The oldest concerns the *Schliffgrenze*[274] on steep or vertical faces of suitable rocks around nunataks, high lands, valley sides and spurs between cirques, i.e. where the ice was active longest and into the latest stages. De Saussure[275] observed in the Alps that rounded forms (*têtes arrondies*) lay below and raw shapes (*cimes sourcilleuses*) above a certain line. Like Hugi,[276] he failed to appreciate the significance of his observation: de Saussure gave no explanation and Hugi, who made his observation about the Aar Glacier, related the contrast to a difference in the nature of the local granite. The true explanation was reserved for later glacialists,[277] though the suggestion was erroneously made that the line marked where the ice was permanently frozen to the ground[278] and therefore gave a minimum altitude.

The related feature, the "erratic height", also noted by de Saussure[279] and afterwards by others,[280] is often connected with boulders of local rocks since

the tributaries occupied the outer strip in composite glaciers.[281] It is frequently distinct in the angle between valleys or at the ends of spurs and, like the *Schliffgrenze*, is absent from firnfields and cirque-basins.[282] Both lines theoretically provide the height and gradient of the ice-flood. In practice, they give a somewhat lower line[283] since the upper ice, being clean and erratic-free, was unable to polish, and frost afterwards removed much of the polished skin at higher levels. The erratics often found no lodgement on the steep slopes or fell to a lower level or scree foot when the ice retired. Many too have been displaced by slips and creep or disintegrated by weathering. Thus the tendency is to raise the limits as more care is given to them. Transfluence across cols or passes furnishes a useful check.

These two lines are conspicuous in the Alps and enable the minimum thickness to be measured fairly easily though there may be stadial lines.[284] Representative estimates in earlier works[285] gave 1000 m over Lago di Garda, 1680 m in the upper Rhône valley, 1160 m at the head of Lake Geneva and 990 m at Geneva itself. Penck and Brückner calculated innumerable figures,[286] the greatest, viz. 2000 m, being near Brig in the Rhône valley and near Meran in the Etsch valley. Since they completed their work in 1909 little has been done on these lines. The thickness, however, has been obtained near Salzach[287] where 1800 m was the maximum and 1500 m was not uncommon, and may be calculated from ice-contoured maps (see p. 701). In the Himalayas depths up to 2000 m have been computed. In New Zealand,[288] the Tasman valley was filled to a depth of 1067 m though the maximum depth in that country was *c.* 2130 m.

In the case of ice-sheets, these flood marks are rare and generally marginal. Among the peripheral "glaciometers" in Europe are Zobten in Silesia and the Polish Mittelgebirge with a thickness of 200 m.[289] North American analogues are Mount Katahdin[290] (5267 ft: *c.* 1604 m)—striae and foreign erratics occur on its highest point—White Mountains, Green Mountains, Adirondacks (5344 ft: *c.* 1627·5 m) and the Catskill Mountains[291] whose summit (4205 ft: *c.* 1282 m) is striated. Scratches, erratics and streaks of boulder-clay indicate a complete burial of Mount Washington[292] (6284 ft: *c.* 1915 m) and at least 5000 ft (1524 m) of ice over the adjoining plains and even greater depths back from the edge. Similar methods gave the Cordilleran ice along the 49th Parallel an average of 800 m and a maximum of 1900 m.[293]

In Scandinavia, where it is uncertain whether or not these signs belong to the maximum glaciation,[294] striae and erratics in the Sognefjord[295] gave 1700–1800 m. The thickness in south Norway was 2000 m, in north Norway 1200–1500 m, and on the ice-axis 800–1000 m.[296] While the ice was deeper in the valleys, the shapes of the Jotunheim peaks are not reconcilable with a surface above 2200 m even at the maximum[297]—nunataks in Okstinder gave a surface at *c.* 1800 m, in Torne Träsk between 1800 m and 2000 m and in the interior of the Ofotenfjord 1000–1200 m.[298] Tanner[299] arrived at a figure of 3570 m and thicknesses of 3000–3500 m have been given for Sweden.[300]

The objection sometimes urged[301] that subsequent erosion, by lowering the plains and valleys, exaggerated the ice-depth is refuted by the freshness of the striae and erratics which, as in the Alps,[302] are clearly of the last glaciation. Later erosion has been negligible (see ch. XXV).

Eccentric icesheds (see p. 667) and the overriding of main water-partings

provide another mode of attack[303]; this has been used repeatedly for Scandinavia. Granite erratics from east Jämtland, for example, were lifted 1000 m on to the top of Åreskutan (1420 m) in west Jämtland, the ice being scarcely deflected as it crossed the cone of the mountain. This implies a thickness in substantial excess of the ridge according to Agassiz's empirical law,[304] founded on modern Alpine observation. The depth on the Baltoscandian ice-divide has been variously estimated at 1000 m,[305] 2000 m,[306] 3000 m[307] or 4000 m[308]; in east Jämtland it may have been 600 m as recently as 11,000 years ago.[309]

We may also infer great depths from the following facts: the Baltic basin did not deflect the ice during the peak of the ice-flood (see p. 710); Baltic flints were raised to *c.* 400 m on the flanks of the German Mittelgebirge—this demands a thickness over the south Baltic of at least twice that amount (see above) or of 1000–1500 m[310]; and although a very small surface-slope is necessary to induce flow, the ice traversed vast distances—an inclination of 1 in 800 from Scandinavia to south Russia would make the central depth roughly 4000 m. In this connexion it may be mentioned that a basal gradient of 1 in 528 has been deemed necessary to induce a flow in ice-sheets[311] and the average slope from Scandinavia to central Germany is estimated to have been 1 in 100.[312] Isostatic depression (see ch. XLVI) must also be taken into account—the North American continent under the heart of the ice may have been 800 m lower than now.[313]

Nansen's assertion that an ice-sheet conforms in section to a regular geometrical curve[314] has often been examined.[315] The curve rises steeply from the edge and then ever more gently. Its ideal shape is not attained[316] in either Greenland or the Antarctic since the form is insufficiently regular and is influenced too much by the buried relief. Comparison with the hypsographic curve of Iceland and Norway shows that it may not comply with any mathematical form.[317] Its immediate determinant factors are not known. It results no doubt from purely glacial laws and is influenced by precipitation and ablation, by the obstruction of the coastal mountains, by the plasticity and physical properties of ice and by the erosion by anticyclonic winds.[318] The Scandinavian ice-sheet was asymmetrical; on the west its surface fell at the rate of 5‰, on the east of *c.* 1‰.[319]

While Nansen regarded the *type de profil de l'inlandsis* (A. de Quervain) as an arc of a circle, Meinardus[320] found that between the heights of 800 and 2,700 m it was a paraboloid or a semi-ellipsoid more curved than a viscous mass as worked out by Lord Kelvin.[321] In the accumulation area, the slope may steepen outwards so that at any section it is approximately proportional not only to the velocity but to the radius at that section.[322]

The inclinations of the ice-borders varied with the temperature of the overlying air and the plasticity of the ice, with the relief and existence of obstructions and with the presence or absence of standing waters about the ice and whether such waters were deep enough by their buoyancy to cause calving. The inclinations have been measured in both Europe and North America, usually from the descent of lateral and terminal moraines, less commonly by the *Schliffgrenze* on marginal nunataks and by the marginal drainage.[323] North American gradients in feet per mile so computed include the following[324]: 50–100 in Washington; 320–370 in Wisconsin; 25–30 along the side of the Hudson Glacier; more than 40 from the Catskill Mountains to the ice-edge; 50 in Montana; 30 in New Jersey and adjacent New York;

100–110 in Massachusetts; 250–280 in New Hampshire; and at least 27 south of Mount Katahdin.

The slope probably exceeded 25 ft/mile over eastern North America—it has been estimated at 30 ft[325] (1:176)—and was probably steeper along the arctic segment where the melting was less, as noticed on modern glaciers, e.g. in Alaska,[326] or in the Pleistocene Alps.[327] These figures may be compared with those for Scandinavia[328] (1–4:100) or with those Antevs[329] collected for the Antarctic and Greenland.

The gradient of the Würm ice in the Hamburg area was 1:330 and in the Ilm valley was at least 1:200.[330]

Calculations for Pleistocene Switzerland were made by A. Favre. Later, Penck and Brückner[331] found gradients of 50‰ and even of 80‰, as on the Piave Glacier, and obtained an average of 12‰ for the subalpine glacier fans. Penck[332] gave similar figures for the Bavarian valleys between the Lech and Isar as did Heim[333] for the Alps generally. In the Allgäu the slopes averaged 25–50‰.[334] The gradient was lessened where mountain barriers stood in the path.[335]

Although the fall did not continue at this rate but became almost zero over much of the interior, the marginal gradient, together with the testimony of the "glaciometers" and the uplifted erratics, as south and west of the Canadian ice-centres (see p. 370), imply that the central ice was enormously deep. Early workers gave figures which critical research has fully confirmed. The following may serve as typical—Antevs[336] has cited others: in the centre of the North American ice-sheet,[337] 10,000–15,000 ft (3050–4575 m) or even 18,000 ft (5500 m) or 20,000 ft (6100 m); near the margin,[338] as in New England and about the Adirondacks, 6000 ft (1830 m); in Montana, 4450 ft (*c.* 1360 m); and over Mount Desert Island, Maine, several thousand feet. Estimates for the Cordilleras[339] include 4000–6000 ft (1220–1830 m); in Puget Sound, 3000 ft (915 m); in Dease Lake valley, 4500 ft (*c.* 1370 m); and over Alaska,[340] 5000 or 6000 ft (1525 or 1830 m).

Estimates for the Scandinavian ice-sheet were of the same order (see p. 42). Over the British Isles the ice was never so thick that the mountains ceased to influence its flow. In the Scottish fjords, the depth was 4000–5000 ft[341] (1220–1525 m); at the head of Moray Firth at least 4480 ft[342] (1365 m); over the Minch, 2700 ft (*c.* 820 m); and in the Outer Hebrides, where the summits of North Harris (e.g. Clisham, 2622 ft; 800 m) and South Uist were unglaciated, 1600–2000 ft[343] (*c.* 490–610 m). It approached 3000 ft (915 m) above the Scottish lowlands[344] and in Skye (the highest peaks of which were completely covered[345]), as well as over the Irish Sea and North Sea[346] and around the Isle of Man[347] where Snaefell was scored along the contour. The ice flanked the west Pennine Chain at 1750 ft[348] (533 m) and passed with a depth of 1700 ft (*c.* 520) and 1000 ft (305 m) respectively over Anglesey[349] and Pembrokeshire.[350] Some of the highest summits in the Mourne Mountains were nunataks,[351] except possibly at maximum glaciation, and the surface during the last glaciation fell from here to the Wicklow Mountains at the rate of 3‰.[352]

The thickness may also be computed from the compression of the drifts and underlying clays by the weight of the vanished ice-sheet[353]—the Miocene clays of Hamburg gave a thickness of *c.* 350 m—and, in the opinion of some, from the amount of isostatic uplift which followed the ice-unloading. Allowing for the uplift that preceded the freeing of the central areas from ice (see

p. 1327), it agrees reasonably well with figures otherwise obtained and amounts to 840 m, 1000 m, 2100 m or 3570 m for Scandinavia[354]; to 1000 m for Iceland[355]; to approximately 1100 m[356] for Spitsbergen and Franz Josef Land; and an addition of between 280 and 695 m to the present thickness in Greenland.[357] The Keewatin ice was thinner than the Labradorean.[358]

The average thickness of the modern Antarctic ice, as given by the abnormally low continental shelf (see p. 1346), is *c.* 550 m.[359]

Ice averaging *c.* 4 miles thick (6·5 km) has been invoked for the great lowering of sea-level that permitted the cutting of the submarine canyons (see p. 1246). The more reasonable figure of 2103 m has been calculated from the lowering of sea-level as measured at the Golden Gate, California.[360]

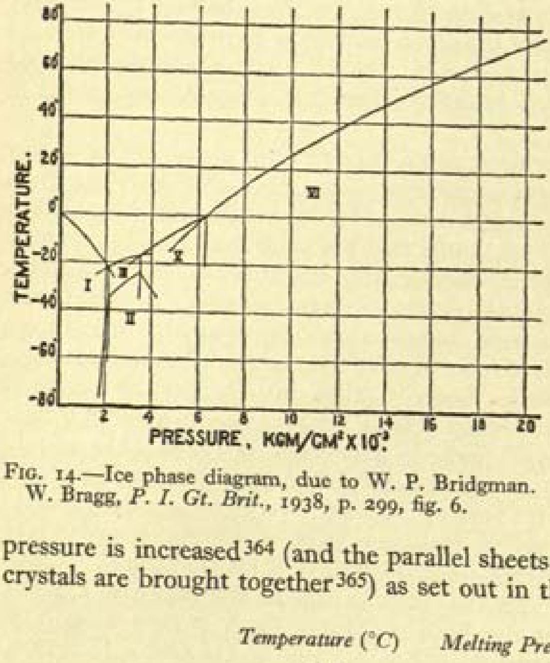


FIG. 14.—Ice phase diagram, due to W. P. Bridgman. W. Bragg, *P. I. Gt. Brit.*, 1938, p. 299, fig. 6.

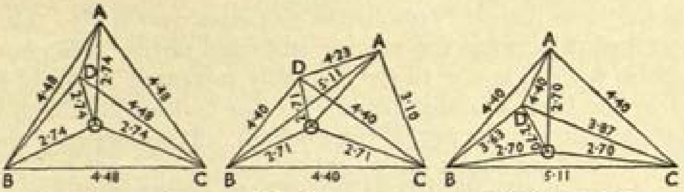
Although Croll's estimates of Pleistocene thicknesses, e.g. 12 miles (*c.* 20 km) for the Antarctic[361] are inordinate, 3000 m or more must be judged probable. The contention that the depth never exceeded a certain limit[362] (up to 716 m or *c.* 3 km) because the ice would melt at the base is not justified. It ignores the ratio of loss by basal melting to surface accretion and the latent heat of fusion of ice.[363] It also neglects polymorphy and the raising of the freezing point through several inversion points as pressure is increased[364] (and the parallel sheets of puckered hexagons of the crystals are brought together[365]) as set out in the following table:

Temperature (°C)	*Melting Pressure kg/sq. cm*
−5·7	678
−10·7	1225
−15·7	1681
−21·7	2070

Several different forms of ice can be recognised and a complete phase diagram can be drawn[366] (fig. 14). In ice II (sp. gr. 1·105) the puckered sheets remain practically unchanged but are forced together by a bending of the bands between them. In ice III (sp. gr. 1·21) the sheets give way. The arrangement of the oxygen tetrahedra is shown in fig. 15[367] (after McFarlan).

Nevertheless plasticity may limit depth[368] though limitation probably results from other causes, such as horizontal flow, surface-melting, wind-erosion and reduced precipitation at high altitudes[369] (see p. 667), especially above the zone of maximum precipitation (see p. 654)—ice may apparently attain a thickness of 10 miles (16 km) before reaching the critical pressure.[370] The Pleistocene ice-sheets may have had a maximum thickness of about 3000 m,[371] a figure which significantly enough is roughly that of the surface of the

modern ice-sheets of Greenland and Antarctica, though these differ greatly in areal extent. Considerations of the earth's thermal gradient have suggested a figure of 1000 m.[372]



Density=0.96 Density=1.21 Density=1.105

FIG. 15.—The oxygen tetrahedra in the three forms of ice, drawn in accordance with the results of McFarlan. In each case O, lying at the centre of the tetrahedron, is equidistant from the four corners A, B, C, D. W. Bragg, *P. I. Gt. Brit.* 1938, p. 301, fig. 8.

Crevasses. Surface-moraines, ablation features and the degree of crevassing determine what Agassiz called the glacier's physiognomy. Crevasses, which occur where tension exceeds the cohesion of the ice, are rare in the névé, especially in its lower part, since movement is slow and changes of slope are few. Nevertheless, they are found towards its edge, where they may be so numerous as to build séracs, and above irregularities in the floor: they are well displayed in maps[373] of the Hintereisferner, Schlegeisferner and Gepatschferner.

Firn crevasses are generally distinguished by their great depth—they may be 80 m deep[374]—since the firn is colder and drier and therefore more brittle than the ice of the tongue. They are also relatively broad (up to 30 m), since surface snows have only a small cohesion. Their upper edges project or are arched over with snow throughout most of the year. They taper to their rounded ends and widen downwards and then narrow as Hugi[375] noticed, the upward narrowing being due to the constant addition of new layers which move less the younger the layer[376] (see p. 100) and leave remnants projecting from the sides. Some crevasses, which are rent open by tension below, stop at discordant layers in the névé and fail to reach the surface.[377]

Firn crevasses occur especially near the steeper sides and at changes of slope. The most important of them carries the specific name of *Bergschrund* (Fr. *la rimaye*) in the Bernese Alps[378] (pl. IIA, p. 33). It follows the semicircular cliff a little out upon the body of the glacier and frequently continues for long distances (it may run for a few thousand metres or cross a névé's whole width), obstructing the climb to the rocks above. Closing in at the bottom, it ranges from narrow cracks to chasms, say, 25 m wide and 45 m deep: a series of crevasses penetrating the névé obliquely may replace it if the rock-face behind is less steep.[379]

The bergschrund originates at the surface some distance out from the rock-wall—the chasm caused by melting between rock and névé is the *Randkluft*[380] (border crack which often forms late in the melt season and develops from the bergschrund)—or where the rock is so steep that snow cannot cling to it, i.e. slopes more than 50°.[381] It approaches and finally meets the rock at a greater or lesser depth, though some authorities believe it is wholly in the névé.[382] In winter, it fills with avalanched snow or is spanned by snow-bridges and hung with gigantic icicles suspended from the upper side. If the bridges are continuous, its presence is indicated by a line of depressions

(Fr. *caveaux*). It opens in spring when the firn sags away from the steeper, upper snows which are thin and frozen to the rock[383] and virtually fixed. They are, however, subject to a slow creep which causes the upper lip to advance slowly, involving a new fissure higher up, until the ice between the two cracks falls into the old bergschrund or passes over it.[384] The position of the bergschrund between the moving and fixed névés (analagous possibly to the landslip fissure) is due to the suddenly increased depth of the snow, controlled by the basin's configuration,[385] or to faulting resulting from tensions and subsidences in the névé[386] (see below). It is obviously absent from plateau glaciers and ice-sheets (cf. p. 47).

Galleries driven during the World War[387] of 1914–18 confirm an observation made previously on the Marmolata Glacier[388] and later on the Hardangerjökull,[389] namely, that the bergschrund has frequently a series of cross floors, 0·5 m apart and 20–30 cm thick, which are rough below and smoothed by freezing waters above. The floors, which extend from side to side and may represent annual avalanche falls across the chasm, are often faulted by glacier movements which here are liable to be sudden and violent.

The tongues have crevasses which are bottom, marginal, longitudinal and transverse. Bottom crevasses (Ger. *Grundspalten*), which have seldom been seen, occur mainly in thin ice near the margin.[390] Marginal crevasses are due to the friction of the sides or of nunataks, e.g. Gaussberg, to the more rapid flow along the centre line, and occasionly to rotation.[391] They point obliquely up the glacier from the edge, usually at 45°, but actually at an angle which varies with the glacier's thickness and velocity—crevasses inclined at increasing angles to the side of the glacier replace them in depth.[392] Longitudinal compressive stress, caused possibly by ablation or a concave bed, swings the direction of maximum tensile stress more transverse to the line of flow so that the crevasses make an angle of less than 45° while a longitudinal tensile stress, resulting possibly from accumulation or a convex bed, increases the angle to above 45°.[393]

Cracks widen into crevasses by movement and by melting of the walls. The opening of a gaping crevasse in a slowly moving glacier is accompanied by an outrush of air, previously imprisoned during glacification by the closing of earlier crevasses and by compressing the chambers excavated by running water. It is the work of weeks or seasons and may be accompanied by lateral displacement and gliding.[394] Whether it cuts across or runs along the margins of the granules is undecided: Hugi[395] and others have thought the former but H. Crammer found the fracture was independent of the granules in ice below 0°C but skirted their margins above 0°C. Hugi's view has recently been corroborated.[396]

Crevasses are arranged fanwise if a glacier abuts against a lateral knob or overrides a buried spur and develops thrust planes.[397] They arise too if affluents encounter a trunk glacier, as on the Petermann Glacier,[398] or where outlet glaciers from South Victoria Land impinge upon Ross Barrier (see p. 169).

Longitudinal or radial crevasses, more or less parallel with the flow, occur if there is spreading or tension, as at the end of a glacier especially of the expanded-foot type, or below a constriction where flow quickens abruptly.[399] They link themselves up the glacier with the lowest members of the marginal series.

Transverse crevasses open if the ice is stretched in passing over an

irregularity in its bed.[400] Maps of the Hintereisferner or Vernagtferner clearly bring out this relation[401]; for a steepening of only 2–3° is sufficient to open powerful crevasses.[402] Such crevasses which curve downstream because of the more rapid flow on the axial line rarely cross a glacier though they frequently join up with the marginal crevasses. If the declivity is both steep and unequal, they build a network of fissures or form splintered peaks, Charpentier's *aiguilles*[403] or de Saussure's *séracs*,[404] so named because they resemble certain cubic cheeses of Savoy. Séracs may, however, arise from pressure.[405]

At the concavity below a step, tension gives way to compression and regelation closes and seals the crevasses; traces are slight when melting has bevelled their edges and closure gives a wavy surface. During the healing process, imprisoned air is expelled in a "sighing" of the glacier and water is driven out in geyser-like fountains.

Plateau glaciers with little or no inclination have few and poor crevasses. Crevasses are absent too from the interiors of Greenland and Antarctica[406] but characterise ice-falls or "dimples",[407] R. E. Peary's "basins of exudation"[408] between the nunataks, in front of which they pass laterally into pressure waves. The gravity stream in the outlet glacier sucks away the converging ice more quickly than hydrostatic pressure can supply it from behind. The wide, innermost crevasse has been interpreted as a bergschrund.[409]

Bosses of rock rising in the bed may raise corresponding "ice-domes"[410] at the surface. These "unborn nunataks",[411] unbroken, shattered or crevassed, are known[412] on the Cornell and other glaciers in Greenland, in the Yakutat Bay region, on Mount Hood in North America and in Antarctica. They are rimmed sometimes by ice-crevasses or have linear debris trains in the lee. The low swells occasionally noticed[413] in the Pamirs and west Turkestan and towards the margin of the Greenland ice (see p. 39) are analogous, as seismic work in Greenland shows. Similar investigations reveal a roche moutonnée projecting 50 m into the sole of the Pasterze and its reflection in a surface undulation 300 m above.[414]

"Glacier canals", found by A. E. Nordenskiöld[415] in North-East Land and afterwards by the Oxford Expedition of 1924[416] (E. Greenland has similar chasms[417]), are parallel, steep-walled rifts in the ice, up to 9 m deep and 30 m broad. They may be due to faulting of the ice (Nordenskiöld) or of the subjacent rock[418] but more probably to tension in the upper ice as it creeps over steep slopes.[419]

The depth to which crevasses penetrate is not certainly known: it doubtless varies from place to place on a single glacier. Soundings, cliff sections in tidal glaciers, and water standing in crevasses[420] or in cylindrical cavities made by closing crevasses[421] demonstrate, however, that few crevasses reach the base. Countless numbers are confined to the upper 5–20 m, i.e. to the upper rigid zone in which movement is by discontinuous fracture and by shearing along discrete planes. Exceptions are situated in the thin margins and at abrupt and high steps in valley floors. Even transverse crevasses which are probably the deepest[422] rarely reach the sole because of compression, though scars which persist to the very end of some glaciers prove that exceptions to this rule exist.[423]

The restriction of crevasses to the upper, frangible layers (where a blow from a ski stick has opened crevasses 45 m long[424] in east Greenland and

Antarctica) may be due to low temperature and small plasticity.[425] Hence crevasses in the cold firn are relatively deep (see p. 45); in Alaskan glaciers they terminate at a definite horizon[426] (which in the South Crillon and Klooch glaciers is shown by seismic methods to be *c.* 30 m) and in Himalayan and equatorial glaciers, which presumably are more plastic,[427] they are fewer than in the Alps. Some cracks may even be due to changes of temperature[428] (Hugi and others thought all crevasses originated in this way[429]), a state of tension induced by flow being a predetermining or predisposing condition.[430] Such cracks, often only as wide as a hair, traverse considerable distances: in the Antarctic, where the abundant crevasses reflect the rigidity and inflexibility of the ice, they have been traced for kilometres.[431]

Banding. Bands of light-coloured porous ice with small grains[432] alternate on a glacier with layers of compact and blue ice lacking air particles. They are a few hundred metres long and 5–7·5 cm or even less than 2·5 cm thick.[433] Their width is usually a few centimetres or decimetres and lessens towards the glacier sides where they merge. They wrinkle the surface by differential melting,[434] as Forbes[435] was the first to notice, and, when vertical, give rise to the more prominent Reid's ridges,[436] the furrows serving as water courses.[437] In Greenland the differential melting is locally reversed and the blue bands sink.[438]

In simple glaciers the bands, each composed of countless lenses[439] (Ger. *Blätter*), are spoon or trough shaped according to the form of the bed.[440] They dip inwards from the flanks and snout as seen in crevasses, ice-caves and Chinese walls, and sweep across the glacier in broad arcs, convex downstream.[441] In glaciers bounded by steep or vertical faces, as in Spitsbergen and Greenland or in the long glaciers of central Asia, the structure (syn. veined, ribboned, laminar structure) stripes the glacier longitudinally.[442] The upturning at the snouts, particularly well seen if the bands are sullied in the line of median moraines,[443] varies from a few degrees to vertical. It is sometimes accompanied by a thickening of the layers, due possibly to exceptional growth from penetrating surface waters.[444]

The structure[445] in narrow valleys, e.g. in the Hochjochferner, remains closed to the end. It is compressed if tributaries enter or valleys are constricted and is less inclined if they widen out. In composite glaciers, it remains distinct and separate and more or less parallel with the planes of contact. It is seen in transverse section as a series of loops, open upwards, which may be very compressed as in Zillertal where each glacier component is narrower than it is deep. In all kinds of glaciers the definition of the zones becomes less sharp in depth and ultimately disappears.[446]

Since Guyot[447] described the structure in 1838 (P. Vidalin[448] had noticed it in the 18th century) it has often been investigated. Agassiz,[449] who had the good fortune to select the simple Lower Aar Glacier, maintained that it was merely the névé stratification pulled or squeezed out. This view, which has the merits of simplicity and probability and contradicts no known fact, has found much favour,[450] especially during the present century. The stratification may be highly distorted or indeed completely obliterated by ice-falls, the opening and sealing of crevasses and complex ice-flow.[451] But if a glacier is undisturbed throughout its course,[452] e.g. the Plattferner or Übergossene Alm, or is inappreciably constricted,[453] as in Norway's plateau glaciers, Kilimanjaro's dome glacier or the Antarctic ice-sheet, the stratifi-

cation is recognisable to the edge. Where the broad névé passes to the narrow tongue, as in the great valley glaciers of Greenland,[454] the original planes are laterally compressed and much modified. Yet the transition from stratification in the névé to banding in the tongue has been followed by H. F. Reid[455] and members of the Third International Conference on Glaciers in 1905[456] on the Forno and Aar glaciers, by Crammer[457] on the Marzell, Vernagtferner and Obersulzbachferner, by Rekstad[458] on the Folgefonn, Jostedalsbrae and Snehaettabrae, and by Øyen[459] on Jotunheim. Crammer in particular has fully described the passage from the undulating stratification of the névé into the synclines and anticlines of the tongue while Hess,[460] experimenting with wax and hydraulic presses, has simulated banding as stratification, reproduced the upturning at the snouts, and shown that each glacier has separate dirt-bands (see below). The parallelism between the periodicities in the thickness of the bands and climatic periodicities strengthens the belief in a descent from stratification.[461]

The readjustments consequent upon granular growth, contrary to expectation, involve no change in the relative position of the individual grains, blue bands and banding continuing unbroken by this process.

Other glacialists,[462] following Tyndall and Huxley,[463] who like Forbes studied banding below the ice-fall on the Mer de Glace de Chamonix, contend that it is allied to slaty cleavage and arises when pressure is strongest, as in constrictions, at margins and confluences, and at the foot of ice-cascades. This view, almost unchallenged until the beginning of this century, asserts that the bands have not descended from stratification since this is wanting in central Greenland[464] or would be destroyed by ice-falls; that they often cut the stratification at high angles,[465] intersect one another[466] and occur in reconstructed glaciers[467]; and that the least disturbed glaciers have the clearest stratification and the best developed bands.

Nevertheless, the banding in reconstructed glaciers is apparently a stratification feature[468] due to successive falls and the pulverisation and melting in the intervals; for the pressure in these glaciers is too small to induce banding. Intersection of stratification and banding may arise from unconformities in the névé[469] or from the stratification's intersection with ice-filled crevasses.[470] These ice-veins which are of varying width, full of air-bubbles and free from dirt, project as ridges below ice-falls[471] because they reflect strongly. They represent old crevasses filled with snow. The latter melts in summer and on freezing forms water-ice (Tyndall's "white ice-seams", Agassiz's *glace d'eau*) which when pure is blue but is frequently stained brown, grey or green with wind-blown dust. The water freezes from the sides and the crystals grow inwards as "candle-ice" to meet in a layer of bubbles, Drygalski's *Mattfläche*. Horizontal columns stretch from wall to wall, though currents of water passing along the crevasses sometimes curve the crystals with them. Pressure subsequently modifies the water-ice into glacier-ice. The more rapid flow of the surface curves and inclines the veins and, if the glacier is long enough, forces them to become nearly horizontal or parallel with the bed. Here they may be orientated with the banding and if filled with dirt may generate dirt-bands (see below). In this way, they may produce true banding,[472] as emphasised by Forbes and early writers as well as by recent glacialists and the International Glacier Conference of 1905, and may intersect the banding inherited from stratification. But banding in general cannot originate in this way; for crevasses, being shallow, disappear by ablation in a

relatively short distance below an ice-fall[473] and many glaciers have no bands below their icefalls.[474]

Banding therefore arises both from stratification ("Alaskan" bands[475]) and from crevasses or glide-planes or foliation[476] ("Forbes's" bands) though it may be truer to say that gliding takes place along the banding.[477] Shear planes, arranged parallel with the flow and the glacier sides, are sealed by freezing waters. Since healed planes of this kind resist shearing better than white layers do, new glide planes tear across where confluence, constriction or widening of glaciers require a new orientation of shear planes. Sets of intersecting shear planes thus originate. That banding may arise in this way is shown by the parallelism of bands and shear planes; by their abundance in places of marked differential movement; and by the steeply inclined banding in the lower parts of some Antarctic bergs which are almost horizontally stratified.[478]

Dirt-bands. Narrow, dark or black bands of denser ice sweep across many glaciers, e.g. Mer de Glace, Glacier du Tacul, in numerous graceful curves or sharp hyperbolas concave towards the firn[479] (pl. IIIA, p. 64). They are the dirt-bands[480] of Forbes and Tyndall (Godeffroy[481] had already noticed them), the *chevrons* of Agassiz,[482] *écailles* of C. Martins,[483] and *Ogiven* or ogives (Gothic pointed arch) of the brothers Schlagintweit.[484] Inconspicuous or prominent (as in the evening light or near lateral moraines), they appear as straight lines below the firnline and become better developed, more curved and more widely spaced down the glacier. At a confluence, they are separate but gradually merge to become meer hooks on the curves or be finally eliminated.

Dirt-bands may be the outcrops of dirt along the stratification,[485] as is indicated by their regularity and coincidence with blue bands, their lack on the Bosson Glacier where the firn mantle sweeps up and over the summit of Mont Blanc[486] (as "dirt zones" they may represent the melting not of one but of several years[487]), and by the pollen layers (see below). They may also be the outcrop of banding due to glide planes[488] or owe their dirt to penetration from the surface[489] or from the base[490] or to enclosure in sealed-up crevasses.[491] R. Streiff-Becker[492] has recently associated them with pressure waves below ice-falls where the ice meets with some resistance. The intervals between the bands may mark the annual progress of the ice,[493] as Tyndall conjectured from observations on the Mer de Glace and Forbes had done 17 years earlier, while their number may give a glacier's age[494]: a glacieret in the Urals Mountains had 220 such annual layers.[495] In the East Twin Glacier of Alaska they have been attributed to the periodic occurrence of obstructed extrusion flow at the base of an ice-fall.[496]

Pollen determinations[497] in the Swiss and Tyrol Alps—up to 5000 pollen grains per cubic decimetre of ice occur—show that in the superficial and central parts of a glacier banding and dirt-bands are coincident with stratification but beneath and at the margins they are due to slipping and structural movement.

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Chapter III

ABLATION

Effect on grains. Névé and glacier grains are pure ice and are associated with intergranular waters which have fine rock powder and salts[1] which derive from the original snow. The difference between land- and sea-ice is in this respect one of degree: sea-ice encloses much brine, land-ice merely a trace. The presence of the saline skin in glacier-ice was suggested by the freezing temperatures of saline solutions of varied concentration and by studies of the coefficient of thermal expansion of the ice obtained in these conditions and at different temperatures. It has been proved experimentally[2], the skin having an estimated thickness, which varies according to the pressure, of 0·035–0·007 mm. The ice lattice has an extremely low tolerance for molecular impurities—these are either excluded by the growing crystal or included as bubbles (gas) or globules (liquid).

When the temperature falls, the intergranular salt solution becomes more and more concentrated until finally at $-22°C$, when the solution is 23·5% NaCl and 76·5% water, the cryohydric point is reached. The eutectic, then formed, consists of water, salts and dissolved gases[3] ($NaCl$, NH_4NO_3, NH_3NO_2, CO_2, O_2, N_2, H_2O_2).

Since this temperature is well below that of the glacier the eutectic point is probably never attained and the salt persists as liquid films notably towards the snout. Its higher density, however, tends to drain it away to lower levels to thicken the films at greater depths. With extreme cold the grains freeze firmly together and the ice breaks conchoidally.

Since impurity depresses the melting point, preferential melting will mark intergranular surfaces; experiments show that the intergranular film has a lower melting point than the grains[4]—the preferential intergranular melting is also attributed to the non-crystalline state of the ice at the intergranular boundaries.[5] Water so released arrests the sun's rays and exposes to their action an active surface in the interior far exceeding the outer surface. Gutters separate the grains and drain off the water leaving air spaces which cause the discreet masses of ice to be discontinuous and produce a white coating 2 cm thick. Internal melting, once begun, opens up a maze of delicate capillaries[6] (they are not delineated at low temperatures) so that the snout on a warm day becomes nodular and its grains etch out and finally disarticulate as Hugi noted in 1822—this is the "Hugi effect". In west Greenland the granules are sometimes so numerous that glacier streams sweep them away to build cones fronting the glaciers.[7] In the hollows of the rough capillary surface mud collects and forms small, rounded masses or "mud-balls".[8]

There is also melting along the laminae. Each grain consists of countless flexible thin plates,[9] 0·25–0·5 mm thick, which are composed of numerous parallel and delicate needles[10] and are arranged parallel with the base of the crystal (they are said to be absent from typical Antarctic ice[11]). Drygalski[12] regarded the plates as fundamental and the grains as a product of their union.

The plates emerge on the outside of the grain as fine ridges, noticed and figured by Agassiz.[13] These "Forel's stripes"[14], which may be crossed at right angles by a striping made by melt-channels,[15] have been erroneously regarded as due to two faces of crystallisation alternating[16] or as a peculiar kind of melt-water feature unrelated to the axes.[17] Somewhat different stripes (Ger. *Plattenstreifung*[18]) have been seen on lake- and fjord-ice.[19]

Interlaminar melting is induced[20] by salt-solutions or dust particles and gives rise to Tyndall's "melt-figures"[21] ("water-flowers"; "ice-flowers"). Agassiz[22] figured them and interpreted them as flattened air-bubbles. They appear in early stages as cavities or negative crystals, more or less circular, and later become six-rayed. The rays have rounded terminations, the rounded shape being apparently more general in glacier-ice, the star-shape in lake-ice.[23]

The cavities, which are really etch figures, result from contraction at melting and the star form from the different rate of heat conduction along the crystallographic axes.[24] Since they are perpendicular to the optic axis[25] or in the plane of the laminae, they provide a useful means of finding the orientation of the grains.

Thumb-print ice. The major forms of melting on a glacier resemble the solution features of the karst.[26] Porosity, solubility and planes of weakness, characters of both ice and limestone, permit cone-shaped hollows, water-holes or glacier-mills, caves and subsurface stream courses to form. They facilitate too a corrugation of the surface by long ridges and gullies, more or less sharp edged. These features are found in various parts of the world,[27] particularly on glaciers ("karst glaciers") which are gently inclined and move slowly. Thumb-print ice, meridian holes and craters, with or without a pipe, *Karren*, caves and subsurface river courses and blind and dry surface river channels—all have their counterparts in limestone country. Any difference there may be is due to flow and morainic cover in the case of ice.

Thumb-print ice[28] (Ger. *Firnschale*), which occurs on the surface of the ice as well as in crevasses and ice-caves, resembles solution hollows in limestones. It has irregular, saucer- or funnel-shaped depressions, *c.* 10 cm in diameter, and originates, especially at the end of summer, by solution, insolation, warm winds or intercrossing systems of wind ripples. H. Spethmann[29] distinguished in Iceland three modes of formation, the first due to warm winds, the second and third being modifications respectively of wind ripple marks and of the thin cover of dust distributed as ripples. The prime cause is some irregularity in the density of the snow. Though compared with *nieves penitentes*[30] (see below) and regarded as their incipient forms, thumb-print ice is seemingly quite distinct.

Penitents. *Nieves penitentes*, so called because they stand like hooded monks in the Spanish processions of Holy Week or as devotees wrapped in shroud-like robes doing penance (Fr. *nieves de pénitentes*; Span. *las nievas penitentes*; Ger. *Büsserschnee, Kerzenfelder*), were first observed by Darwin in the Cordillera of Mendoza.[31] These most fantastically beautiful and interesting of all forms of snow are chiefly found in South America between latitudes 24° and 36° S.,[32] as in Argentina, Chile and Ecuador, and were therefore deemed peculiar to the Andes. They occur, however, elsewhere between 40° N. and 40° S. Lat.,[33] as on Kilimanjaro, in Tibet and the Himalaya and Karakoram Mountains,[34] and in the Canary Islands,[35] while

rare and related but rudimentary types, differing possibly only quantitatively, are found in the cold temperate zones,[36] e.g. the Alps, West Prussia, the Brocken of the Harz, Black Forest, Bulgaria, Iceland, Mount Etna, the Sierra Nevadas of North America, Hawaii,[37] and even in Greenland[38] and Antarctica.[39] Their lower limit follows a similar course to that of the climatic snowline.

Penitents, in their equality of detail and grouping of the whole, stand out with rare uniformity, symmetry and regularity. They are roughly elliptical and pyramidal and occasionally acicular. The apices lean over equatorwards at an angle which lessens with latitude and corresponds with the elevation of the sun at midday.[40] Their height, which is usually 1·5–2·0 m but may be 6 m or even 7 m,[41] depends apparently upon the season and upon the stage of evolution; it diminishes as the altitude increases, each array of penitents presenting an orderly appearance. They are arranged on horizontal or inclined fields in parallel rows[42] (pl. IIIB, p. 64), never united into continuous ridges, and run east–west in Hawaii, central Africa, the Himalayas and Andes. On Kilimanjaro, two sets of ridges were described by F. Jaeger, F. Klute and C. Uhlig. W. H. Workman's eight types[43] are not all genuine penitents: they include diverse forms such as séracs and decapitated shafts of glacier-tables.

F. E. Matthes[44], who has named the forms at higher altitudes "sun-pits" and the shallow forms at lower elevations "honeycombs" or "sun-cups" (= thumb-print ice), has shown the progressive development of sun-cups, pits, spikes and pinnacles with altitude.

Penitents are almost invariably associated with névé of uniform density and exceptionally cover its entire surface.[45] They are rarely found in avalanches[46] or in glacier-ice[47]—they occur in parasitic glaciers in the Karakoram Mountains[48]—and never wholly in freshly fallen snow, though snow, firn and ice penitents are distinguishable and a penitent from the cusp to the base often grades from soft snow through hard névé to hard and compact ice.[49] If snow falls on a field already well developed, the evolution continues in the new snow, having the original form as a base. Annual and perennial forms can be recognised.

These striking features result from the strong solar radiation acting in clear cold air upon névés which are uneven in surface or texture; their material is denser than that around them.[50] That they require peculiar conditions is fairly certain but what these are must remain somewhat obscure until careful local studies have traced a full developmental series. They are usually attributed to intense and prolonged solar radiation in regions with uniform meteorological factors, a constantly high sun's altitude, summer cloudiness and dryness, and a temperature over long periods below 0°C.[51] The unusual combination of intense solar heat and cold dry air is essential. In Hawaii the development is most effective where the motion of the sun near noon is practically in a vertical plane with only slight movement in the azimuth. Hence they have been regarded as a "tropical glacier type"[52] or "tropical volcano type",[53] arising when melt-water is diffused or evaporated in the firn[54] or from some inherent or antecedent irregularity or unevenness of the surface.[55] This may be due to snow collecting irregularly in calm air or to a density difference produced by wind packing[56] or arrangement of dust by wind,[57] by avalanches or creep,[58] or by partial fusion which contracts the volume.[59] Analogy with limestone *Karren*[60] is probably erroneous since

penitents do not follow the gradient. Though they may have been cut from continuous parallel ridges[61] this seems doubtful.[62]

While wind may not always be essential, since penitents are occasionally found in wind-sheltered places[63] or oblique to the wind (these may be local deflections, by rock walls or valleys), they would seem to have been formed in general under its influence[64] which, by producing an excess of evaporation, co-operated in evolving the fundamental type. They are best developed in the Andes because here the winds have a fixed direction.

Dust wells. Dust scattered over a glacier aids ablation considerably. Its greater specific heat or the action of melt-waters when it is once submerged cause it to melt its way into the ice to form dust wells.[65] In the Antarctic the few dust wells, of a medium diameter, are *c.* 2 cm deep; in Spitsbergen they are 10–30 cm deep and in Greenland 5–10 cm broad and 50–60 cm deep, shallowing inland. The depth, which varies with the nature of the ice, is apparently a function of latitude: it expresses the difference between direct and indirect ablation and is controlled by the inclination of the sun's rays and their absorption in penetrating the water to the sand grains. The wells persist from year to year and build definite horizons in the ice, except in the lower parts of glaciers where direct ablation is greatest.

The water in the wells freezes at night into needles radiating inwards from the walls. By repeated freezing these form "glacier-stars" which completely fill the hollow and later become granular by pressure. The included air-bubbles also radiate and merge in the centre as a bubbly core.

Meridian holes. Meridian holes (*trous méridiens*) or bagnoires (syn. *Mittaglöcher*, F. Keller; *Kellerlöcher*, E. Desor) are the type of dust well which, first correctly interpreted by S. Studer in 1787, deepens to the north and is semicircular, the chord of the arc being directed southwards. These glacier "sun-dials" become shallower with altitude, as on the Rhône Glacier.[66] On the Gorner Glacier where they are extremely numerous they are often 1·5 m across and 0·8–1·0 m deep.[67] Greenland has similar forms.[68]

The slight diathermacy of the ice[69] enables dust and blocks, though completely buried to a depth of 2 cm, to melt the ice around them. The water escapes by the capillary net and a prism of space develops above the object with a corresponding base-section—Agassiz[70] sought to explain erratics on roches moutonnées in this way.

Cryoconite. A. E. Nordenskiöld[71] described as "cryoconite" a fine grey powder which he found strewn in heaps or in dust wells (cryoconite holes) over the ice of Greenland. Subsequent research has discovered it in Spitsbergen[72] and Antarctica,[73] where thaw and melt-water help to develop them. Nordenskiöld, with others,[74] ascribed to it a cosmic origin because it contains spherulites of magnetite. But the cryoconite has been transported by the wind from nunataks or naked rocks and moraines along the ice-margin[75] as Nordenskiöld[76] afterwards admitted for part of it. Thus the dust lessens into the interior of the ice-sheets[77] where it is practically absent—the small extent of the ice-free areas from which it could be swept readily account for its virtual absence from the Antarctic[78]; its composition is similar or identical as analyses prove[79]; and it contains green algae, rotifers, leaves, pollen, insects, small birds and decomposition products of ptarmigan excrement.[80]

Glacier-tables. The influence of detritus on ablation is well seen in the glacier-tables, a name first used by Hugi.[81] As de Saussure and G. Studer

noted,[82] boulders stand on lengthening stalks or pedestals of ice, up to 0·5–1·0 m high[83] in the Alps, which have been protected while the surrounding glacier has wasted away: lithology, colour, shape and position of the boulder are significant. They occur singly or profusely, notably near raised median moraines or the inner edge of lateral moraines,[84] but rarely in badly crevassed areas or the lower parts of glaciers where the air temperature is higher. The zone most favourable to the development of tables moves up the glacier as the summer advances. The protecting blocks slant equatorwards because the melting is unequal, and slide off their shafts in that direction[85] or toward the heavier side: this lateral wandering helps to widen moraines towards the snout. The distance through which a boulder moves each time depends upon the pillar's height, the shape and size of the boulder, and the nature and slope of the surface. If this is steep the rock may slide quickly with jumps of ¼ km and inscribe pronounced furrow-like tracks.

Tabling occurs once a year in the Alps,[86] occasionally twice, though tables may last for 15–20 years.[87] It is rare on névés and avalanches—"snow-tables" do, however, occur[88]—or in polar lands where melting is small and the sun is low, though tables have been seen[89] in Spitsbergen, on the Malaspina and other glaciers of Alaska, and in the Antarctic (Heard Island). The average height is less in high than in low latitudes.[90]

Moraines too protect the ice, as de Saussure[91] observed, so that linear moraines rest on walls of ice. Their relative height is enhanced by the lowering of the surrounding glacier, aided by radiation from the moraines. Median moraines may in this way be margined by depressions[92] or even embedded in the ice.[93] On the Bowdoin Glacier, west Greenland, a moraine was sunk 3–5 m, and longitudinal depressions, up to *c.* 90 m wide, mark the course of some Antarctic moraines.

Thin stones, particularly if dark coloured, behave differently from thick ones. Like leaves or insects,[94] they sink vertically and rest in hollows as J. H. Hottinger[95] first correctly surmised in 1706, though he erroneously thought they thus bored their way to the base. Others retain their positions, since their critical thickness, which varies with climate and latitude and with the purity of the ice,[96] neither facilitates nor retards melting. On the Fedchenko Glacier this thickness was 20–25 cm. Wedge-shaped stones naturally move unevenly. Studies in Wyoming[97] show that blocks which shield the ice at the beginning of the summer melt their way into it as the season advances, especially if they rest in the middle of a glacier.

Craters. Glaciers are sometimes studded with large hollows (Fr. *entonnoirs*; Ger. *Gletschertrichter*). These have been correlated with moulins, with the enlargement of crevasses, with vertical erosion and with undermining and collapse and the swallow-holes of the Karst.[98]

Sand-cones. Sand-cones ("melt-cones", "sand pyramids", "ant heaps"[99]) are related to glacier-tables and have been widely noted[100] since S. Pálsson observed them in Iceland[101] (1772) and G. Studer[102] described and correctly interpreted them in 1783. They arise if much material gathers, as near moraines, above moulins, along dirt-filled crevasses or thrust planes, or on deserted lakes and stream-channels.[103] According to their nature they are denoted dirt, sand or gravel cones. They possess an ice-core and rise to 24 m on the Malaspina Glacier,[104] to 30 m in Iceland,[105] and to 85 m on the Hispar Glacier of the Himalayas[106]: a limit is set by rolling or sliding to the

perimeter where melting compels the cone to spread. A collection of small stones, a single weathered block or the shaft of a glacier-table which has lost its protecting block may produce a like result[107] (pl. IVA, facing p. 65).

Factors governing ablation. Ablation, a term of Agassiz,[108] which connotes the joint processes of melting and evaporation that contribute to the consumption of snow and ice, is superglacial, englacial and subglacial. Surface ablation increases steadily toward the snout[109] as measured by bores, stakes and tabling and seen in the glacier's progressive diminution in cross-section, in the melting out of englacial debris, and in the increasing importance of surface and marginal streams.

The emergence of boulders was regarded by the Swiss peasants and by early writers, e.g. F. J. Biselx, F. J. Hugi, G. Bischof and E. de Beaumont, as due to a kind of organic function, the impurities being expelled by extrusion from crevasses upon freezing[110] or by lateral or forward pressure and dilatation.[111] T. de Charpentier[112] suspected it was due to surface melting, a conclusion later demonstrated[113] by his brother and J. G. Altmann.

Ablation is affected[114] by the sun's rays bearing directly upon the surface or acting indirectly by reflection from rock-walls or by conduction through morainic debris, by warm rain and thaw water, by evaporation and wind, and by calving. Radiation is supreme on tropical glaciers, as on Kilimanjaro,[115] and where its angle of incidence is high. Thus, the ablation on the Vernagtferner, with an angle of 83° to the summer rays, is 20–25% higher than on the Hintereisferner with an angle of 57°.[116] For this reason, ablation diminishes with increasing latitude, though the amount is complicated by the duration of the radiation, the condition of the air, and the strength and temperature of the wind. Since melting depends upon temperature, unless there are surface snows,[117] it will be more in summer and during the day (in the Alps, there is almost none during the night) and on the sunny side. In polar regions, although restricted to the few summer months, it is relatively large on account of the 24-hour day with continuous light, strong radiation and little cloud. It is greater in west than in east Greenland,[118] and is especially high on plateaux whose whole surface is exposed to radiation during the circuit of the sun. Even the highest ice in North-East Land is attacked.[119]

Evaporation which depends upon humidity, wind, pressure and the temperature of the ice-surface is by no means negligible[120] and is important at greater altitudes and in winter.[121] It increases somewhat in proportion to the square of the wind velocity[122] (whence its importance in the Antarctic) and is indirectly proportional to the air pressure.[123] It is high in the tropics[124] and where, as in Greenland and the Antarctic, winds descending upon the ice from the high surrounding walls and suffering compression, are relatively dry. It is affected by insolation[125] which, as in the Antarctic, may be considerable.

Wind is very important[126] in North-East Land and in Antarctica where its exceptional velocity is largely responsible for the present starving of the glaciers. It inflicts loss by evaporation, even where the air is below freezing point, and by mechanical abrasion,[127] though this is insignificant in west Greenland.[128] In the Antarctic (Adélie Land), it sweeps clear a strip 60 km broad.[129] Its action is unusually strong when it has a foehn character,[130] as on Jotunheim in Norway and on the Karajak and other coastal glaciers in west Greenland. Here the foehn which beats in winter and spring on the

sea-ice with great force and tears it and drives it out to sea, may have a temperature as high as 10° or 16°C in winter.[131] In the Alps, the daily ablation may rise from 2 or 2·5 cm to 9 cm.[132]

That rain exerts an influence was shown by Agassiz[133] on the Lower Aar Glacier. Its significance, however, may be overestimated. It is naturally unimportant in polar lands and even in the Alps is only a minor factor because of its low temperature which is 1·1–2·8°C below that of the air.[134]

A glacier may also be arrested by unusually big moraines deposited at an earlier period[135] or by the sea, in the case of a tidal glacier,[136] where as in the Turner Glacier, Alaska, the ice ends in a fairly straight cliff. The strength of the melting and calving depends upon whether the water is fresh or salt, deep or shallow, quiet or actively moving. Rivers which pass at the snout may also successfully attack a glacier's front[137] in the same way as they are known to check the advance of a dune.[138]

Ablation is much diminished if moraines are abundant. They may reduce the daily amount in the Alps[139] by 1 m or to only one-third of that of dirt-free ice. On dirt-protected parts of the Hintereisferner, the reduction is *c.* 15% and on the Mer de Glace in the ratio of 5·3–2·9. On the Alpenerferner, ablation actually lessened appreciably towards the snout owing to protection by median moraines. Hence, excessively dirty glaciers, as in the Himalayas or the Lower Aar and Z'mutt glaciers of Switzerland, are longer than they would otherwise be. Excessive debris at the snout, as in Alaska,[140] may cause stagnation.

Condensation upon the ice, as Hugi and Rendu early recognised, provides further modification.[141] It is only slightly less than the evaporation in Sweden and, as on the Rhône Glacier and at Davos, may exceed the loss.

Since A. Escher sank stakes into the Aletsch Glacier in 1841 to ascertain the rate of ablation and Agassiz[142] made similar observations on the Aar Glacier, very many like determinations have been made,[143] as on the Mer de Glace, Aletsch Glacier, Pasterze, and Jostedalsbrae, and more recently and accurately[144] on the Rhône Glacier, Lower Aar, Vernagtferner, Hintereisferner, Pasterze, and in Greenland, Yakutat Bay region, and in North-East Land where formulae have been deduced relating ablation to meteorological factors.[145] Mean monthly values on the Vatnajökull[146] show that the higher the altitude the more ablation is confined to the summer and that, as is probably true of most glaciers whatever their type, the ablation decreases rapidly with altitude near the margin. Observations on the Rhône Glacier and Hintereisferner suggest that the mean annual ablation below the snow along the length of the glacier varies almost rectilinearly with altitude[147] (see below).

Considerable difficulties attend these investigations; for example, the rods affect the ablation around them and the surface decays unequally, even in places 1 m apart, and the snow also becomes compacted.[148] In recent times, ablation has been measured by recording instruments, viz. the "ablation-gauge"[149] and the ablatograph.[150]

Enough is now known to obtain a fair quantitative idea. Its annual amount at the Alpine firnline is 2–2·5 m.[151] It diminishes with height[152] at the rate of *c.* 1 m for every 100 m on the Rhône Glacier (but varies much from year to year) and at an average of 1·25 m on the Hintereisferner. Greenland[153] registers a like diminution, the annual loss at 360 m being 2·05 m and at 675 m 1·05 m. Other Greenland figures, of probably local interest only, are

as follows: 50 m, 4·4 m; 270 m, 3·9 m; 570 m, 3·6 m; and 950 m, 1·6 m. Hess has given the annual loss 400 m below the snowline at *c.* 4 m in the Alps, *c.* 2 m in west Greenland, *c.* 3·3 m in Scandinavia and *c.* 1·4 m in north-west Spitsbergen. That these figures represent enormous losses may be gained from the following estimates[154]: one-fourteenth of the ice of the Col de Géant Glacier melts in its passage from the source to the snout; the loss on the Mer de Glacier de Chamonix was 50 m in 15 years or one-eighth of the glacier in 50 years; the Vernagtferner lost in 40 years 239 million cu. m and the Hintereisferner 10·2 million cu. m per annum (see p. 141).

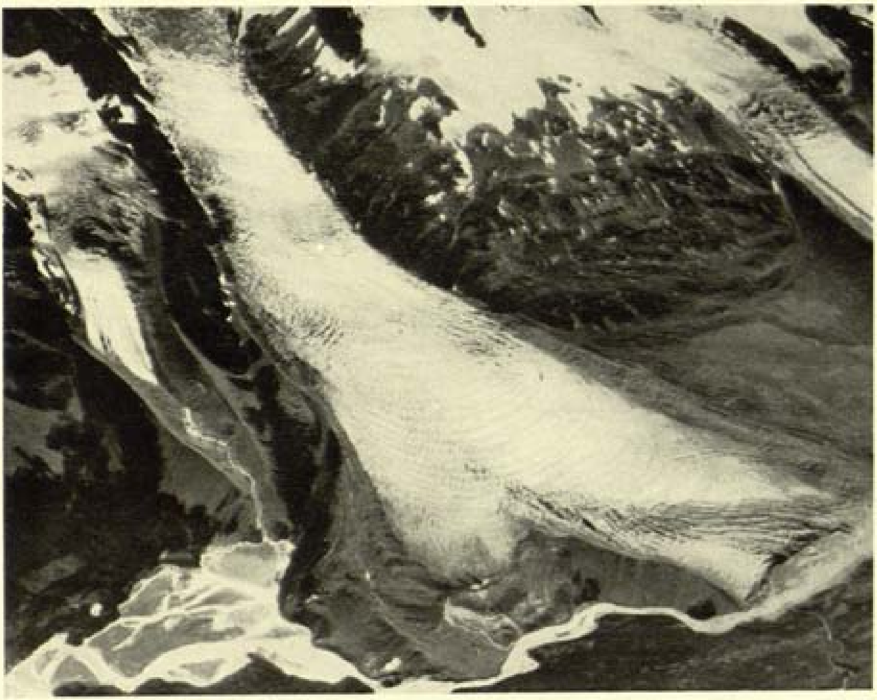
Superglacial streams. In subpolar and high polar glaciers little or no water runs off: it freezes and crystallises on the glacier. In temperate latitudes melt-waters do occur. They gather into rills and streams which, notably during spring and summer, course along radiation gullies margining the ice or channel it in meandering canyons, up to 10 m deep, with specific fluviatile forms[155] (pl. IVB, p. 65). Their velocity is high because their sides are smooth and they are generally clear and free from debris.[156] The gorges are narrower than if cut in rock since ice-flow constantly replenishes their walls. They are 1 m deep in small Alpine glaciers but become deep and broad, with lake-like expansions if the ice has low slopes and few crevasses, as on the Malaspina Glacier and the marginal ice of Greenland. Extreme cold makes them shallow in the Antarctic.[157]

Because of the serious downward percolation which takes place even at high altitudes, rills are rare in the névé except at the steep periphery,[158] where, at least in certain cases, more water may be received by the glacier than anywhere else over its whole surface. Shallow ponds and lakes are found above the snowline, especially in the lee of nunataks. Surface streams which vary in volume according to the season, time of day, and weather conditions,[159] are few above the snouts owing to percolation through the capillary net and crevasses.

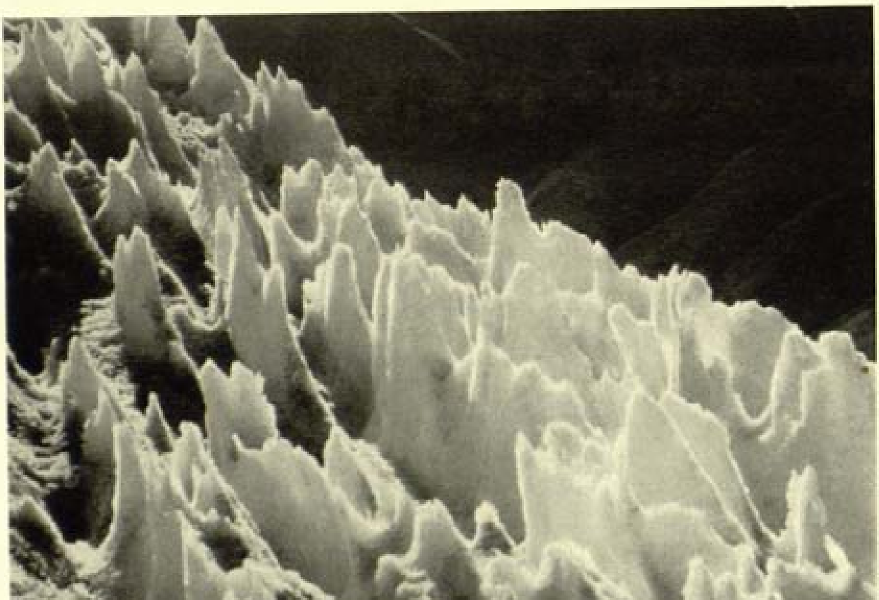
Surface streams and crevasses, especially the transverse type, are naturally mutually exclusive. Streams plunge down crevasses and vertical cylindrical shafts or moulins to join the englacial and subglacial drainage. In little streams, these "glacier-mills"[160] (Ger. *Mühle*) may be dry during the winter; in big ones, only a fraction is swallowed up, the rest flows on. The shape of the shaft,[161] which is perhaps less regular than commonly supposed, is determined by ice-pressure and by melting furthered by the descending water or ascending air. The diameter is often only 1–2 m but may be as much as 6 m: the depth is considerable though mostly unknown (see p. 47).

Moulins are found in almost all glacier-regions, as in Greenland.[162] They are parallel with the ice-flow[163] and move with the ice around them except towards its margin where they tend to be stationary. Lines of deserted moulins, serving frequently as blow holes for escaping air (Ger. *Windlöcher*), pierce the surface below ice-falls (Tyndall observed six of them on the Mer de Glace) and persist sometimes for several months. They disappear, particularly towards their base, as the ice closes in around them and in their upper parts by surface ablation. Ultimately there remains only a blue circular patch, the cross-section of the cylindrical plug of frozen ice, which usually lies on a crevasse scar[164] (see p. 47).

Basal melting. Loss by basal melting is difficult to assess. While some think it exceeds surface-melting[165] it may be trifling[166] in accord with



A. Ogives and crevasses on small glacier east of Barley Lake, St. Elias Range, Alaska [H. Bradford Washburn]



B. Penitents on the snowfields, Barrose region, north-west of Maipó [C. Troll]

A. Dirt cones, Kverkfjöll, Iceland [C. W. M. Swithinbank]

B. Superglacial stream, glacier east of Arkhangel Bay, Novaya Zemlya [O. T. Grönlie]

(facing p. 65)

observations in west Greenland, with calculations on the Hintereisferner, and with theoretical considerations which make it only 2% of the total melting, or, as on the Rhône Glacier, 4·3% of the surface melting. The factors involved are earth's heat, springs, basal and internal friction, and especially waters and air-currents which have penetrated from the surface, particularly near the margin. The earth's heat is unquestionably a factor[167]—it has been estimated at $2{\cdot}5 \times 10^{-6}$ cal/sq. cm/sec[168]—and unlike the other factors is constant in its action. Its effect, however, is hard to evaluate; for subglacial temperatures have, generally speaking, not been measured. Such melting is probably very subordinate if glaciers lie high and have little gradient, e.g. Lower and Upper Aar Glacier, but is important[169] in glaciers which run in narrow gorges and at low levels, e.g. Lower Grindelwald Glacier. É. de Beaumont[170] calculated that it melted an annual thickness of ice of 6·5 mm. Under the Rhône Glacier it was less than 2·8 l./sec/sq. km.[171] The basal melting of Alpine glaciers is less than 5·3 cm/annum.[172] A. Hamberg's[173] figure of 0·006 mm for the daily loss is in rough agreement with recent calculations which, though varying a little among themselves,[174] justify a working figure of approximately 7·4 mm/annum. The melting by internal friction may be ten times as great.[175]

The Icelandic Jökullhlaup[176] (Icel. *jøkelløb*) is exceptional in this respect. These devastating floods have been recorded from A.D. 1201 onwards and 17 are known from the last two centuries, including the *débâcles* of 1721 and 1755—the most celebrated localities are Katla at Myrdalsjökull and Grimsvötn in Vatnajökull (see fig. 84, p. 454). They result, as Pálsson[177] noticed in 1794, from the sudden drainage of ice-dammed lakes (glacilimnogene) or the catastrophic melting by subglacial volcanic activity (vulcanogene). The water causes the glacier to swell and the ice to "run out", as the Icelanders call it. It excavates tunnels and when freed makes its way through and over the ice-masses below or sinks down upon the glowing lava, being converted into steam with tremendous explosions. The mixture of water, lumps of ice, volcanic debris and gravel rushes out to spread over the country outside the ice-margin. These jökullhlaups, among the most violent of natural catastrophes, have devastated the coastal plains of south Iceland and have stamped this part of the island both morphologically and economo-geographically. Time after time they have ravaged the coastal areas south of the glaciers, the volume of water thrown out during one short eruption amounting to some milliards of cubic metres: in three days the Jökullhlaup of 1934 gave 64,000 cu. m/sec (the Amazon gives 74,000 cu. m/sec) with a total volume of 10–15 cu. km and that of Katla in 1918 probably not less than 200,000 cu. m/sec.[178] Outwash plains are flooded, bergs 20–30 m high are propelled towards the sea, and huge waves are generated along hundreds of miles of coast.

Vulcanogene "hlaups" were apparently responsible for Iceland's "palagonite formation",[179] a great series, *c.* 1000 m thick, of globular basalts and tuffs, breccias and glacial moraines and fluvioglacial materials, since rocks wholly similar to them have been formed during recent eruptions.[180] Observations on Grimsvötn seem to show that the eruptive products and the course of the eruption change in character when the subglacial eruption becomes a subaerial eruption owing to the penetration of the overlying ice. Similar floods have also been seen on Cotopaxi and in Kamchatka[181] and Spitsbergen[182] and may have been associated with Pleistocene glaciers in the Caucasus,[183]

North American Cordillera[184]—in British Columbia Pleistocene lavas entering glacier-lakes had concave fronts and other abnormal features—and Ross Sea sector, Antarctica.[185]

Subglacial streams are known from their muffled roar which in some large arctic glaciers can be followed for miles. Their undulating tunnels, which rise and fall in response to the irregular bed, appear at the snout as vaults, usually 1–5 m, exceptionally 40 m high[186] (Ger. *Gletschertor*). Small or hanging glaciers may have several caves, large glaciers only one or two. They are axial in Alpine valleys but lateral in the Arctic[187] where the surface ice is very cold.

That streams flow under great hydrostatic pressure is evinced by the springs and fountains[188] which spout out on the ice, sometimes accompanied by cones of ice in the radiating fissures, and by the waters[189] which boil up at the sides and snout in the sea or ice-fjords to a height of 3·5–4·5 m and make appreciable currents, recognised by discoloration as in front of the Muir and Malaspina glaciers. The velocity, however, as in the case of the Mer de Glace, is less than in surface streams[190] of the same grade and volume—fluorescein experiments on the Rhône Glacier[191] demonstrate that it is only half—because the channels are narrow and the enclosing walls exert friction.

Englacial streams, evidence of which is seen in crevasses or ice-caves,[192] e.g. in the face of the tide-water glaciers or in the dry tunnels in overturned bergs, or occasionally in concentric crevasses surrounding a circular depression on the ice-surface,[193] are fed from melt-waters which circulate along the capillary nets or cascade down crevasses and moulins—Agassiz[194] thought each moulin had one channel. Their winding tubes which bend sharply at the intersection with crevasses may have a considerable cross-section; the Muir Glacier had one of *c.* 16 sq. m.[195] In deep glaciers, they probably run at or just below the base of the zone of fracture.[196]

Extraglacial drainage. Superglacial drainage is especially characteristic of active glaciers, subglacial drainage of stagnant ice.[197] Waters streaming on, in, under, or alongside a glacier unite as extraglacial streams; some of the world's best-known rivers, e.g. Rhine, Rhône, Aare and Po in Europe, and Ganges and Brahmaputra in Asia, issue almost full-fledged in this way. The flow varies with the size, aspect and steepness of the glacier (the volume per square kilometre of glacier-basin decreases with increasing glaciation[198]), with the surrounding relief and with the time of day and season and with such "accidents" as the closing of subglacial channels and the disappearance of extraglacial lakes. The régime of rivers fed in this way is largely controlled by the glacial melting,[199] for example in north Italy and the Swiss Alps, and not by precipitation. At the sinking of the sun melting virtually ceases. The effect, first noticeable in the névé, is not immediate but suffers a delay which increases with a glacier's length, just as the rise in a glacier stream lags behind a rise of temperature. Thus on the Lower Aar Glacier[200] the minimum is at 10 a.m., the maximum at midnight, though the Rhône Glacier has earlier times in winter. The daily variation on this glacier increases from spring to summer but in larger glaciers may be concealed by variations produced by the weather.

The seasonal variation,[201] which is reflected in lake-levels, e.g. Lago Maggiore, Lago Lugano and Lago di Como,[202] has a lag in time of onset of maximum and minimum, the maximum reacting more quickly. The volume

falls at first rapidly, then more slowly, and finally passes into the winter minimum.[203] Winter flow[204] is almost nil in some small Alpine glaciers, though snow and percolation into the fluvioglacial material among boulders and rocks make accurate measurement difficult. Nevertheless, there is some winter flow under almost all glaciers, even in arctic lands,[205] for example in Greenland and Spitsbergen. Some of this may be due to basal ablation[206] occasioned by mechanical friction or earth's heat, to the friction of the glacier, and the retention factor of the glacier—this depends upon the size of the firn, the gradient and the amount of crevassing—but most if not the whole comes from springs within the glacier-basin,[207] the amount depending upon geological structure. Thus ablation is either absent or insufficient[208]; and glacier-streams are limpid,[209] have a higher temperature in winter,[210] and contain much mineral matter at this season.[211] Hollows at the surface in certain glaciers in the Caucasus and Karakoram Mountains suggest warm springs immediately beneath.[212] The greater regularity of streams fed by glaciers compared with those which are not so fed caused their widespread adaptation for hydraulic works and for irrigation purposes.

Glacier-streams[213] are a little above 0°C when they emerge, as early noticed by de Saussure[214] and G. Bischof,[215] and warm steadily as the distance lengthens and they come in contact with warmer air above and warmer ground below—the rise induced by this factor is 0·47°C per 100 m of fall.[216] The actual rise depends upon the stream's volume and velocity and the shape of its channel. The temperature varies[217] too with the time of day, from day to day, season to season and year to year.

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S.* 250, 1952, 553. **185** *373*, 2. **186** *730*, 244. **187** *30*, 278; *250*, 218. **188** *20*, 252, pl. 16; *415*, 48; *659*, 172; *660*, 44; *1455*, 808; D. M. MacMillan, *B. Am. Mus. N. H.* 56, 1928, 431; H. Schatz, *Z. Gl.* 27, 1941, 141. **189** *1440*, 57; R. Brown, *Q. J.* 26, 1870, 681; G. F. Wright, *A. J. S.* 133, 1887, 8. **190** A. E. Martel, *Spel.* 4, 1898, 158; G. & J. Vallot, *Ann. Obs.* 4, 1900, 19. **191** *748*, 223; *1110*, 103; F. A. Forel, *Spel.* 4, 1898, 156, 163. **192** *1384*, 32; *1679*, 114; *1681*,

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Chapter IV

CLASSIFICATION AND DISTRIBUTION OF LAND-ICE

Classification

The classification of glaciers, until the end of the last century, rested mainly upon those of Norway, Greenland and the Alps. Thus the classification then generally adopted[1] was that of Heim[2] who distinguished an Alpine, a Norwegian and a Greenland type. A fourth type, variously termed Alaskan, Malaspina or Piedmont, was added after the Malaspina piedmont glacier of Alaska was discovered and described (see below).

More recent classifications,[3] which resulted from extensive and intensive studies in polar regions since about the year 1890, recognise the importance of polar ice, and show more complexity and differentiation of types. They are based alternatively upon topographic forms and land relief, upon the intensity of glaciation, or upon the stage of glacial development. Others divide the glaciers according to the area of ice-supply, of ice-movement and of wastage.[4]

Geological rock-types and structures also influence a glacier's shape. Thus porous rocks may explain why the Limestone Alps have few glaciers[5] and the want of ice in Jameson Land, east Greenland.[6] West Greenland too presents glacial contrasts[7]: the gneiss has plateau glaciers, and horizontal strata (sedimentary rocks and lavas), more especially basalts, have valley glaciers. This influence was subsequently discerned in Sverdrup Land,[8] where the ice-cap is confined to the Archaean rocks. In Spitsbergen,[9] three regions, related possibly to unequal absorption of heat, may be distinguished: dissected Heckla Hook country with valley glaciers, sandstone plateaux with "step glaciers", and Pre-Cambrian rocks with ice-caps and plateau glaciers. Glacierisation is less among horizontally bedded sedimentary strata, e.g. soft sandstones, than among crystalline rocks[10] in South Georgia, Graham Land and King Charles Land.

The subjoined classification which omits reference to perennial ice or snow-fields or to intermediate or transitional types acknowledges the fundamental control by the relief and alimentation. The several types in each main category fall generally into a receding hemicycle.

Since each type is repeated wherever the necessary relief and climatic conditions reappear, no one type will be peculiar to a particular country. Consequently, geographical names such as Alpine, Norwegian and Malaspina, are better avoided.

Classification of Land Ice

1. Sheet-ice on elevated lands
   - (*a*) continental ice-sheets
   - (*b*) highland ice-caps
   - (*c*) island ice-caps
   - (*d*) plateau glaciers
   - (*e*) carapaces
2. Mountain glaciers
   - (*a*) reticular glaciers
   - (*b*) dendritic glaciers
   - (*c*) valley glaciers
   - (*d*) cirque glaciers
   - (*e*) hanging glaciers

Classification of Land Ice

3. Lowland glaciers
   (a) piedmont glaciers
   (b) expanded-foot glaciers
   (c) reconstructed glaciers
   (d) ice-slabs
   (e) fringing glaciers

4. Transition ice
   (a) shelf-ice
   (b) floating glacier-tongues
   (c) ice-foot

1. *Sheet-ice on Elevated Lands*

The various kinds of sheet-ice (Drygalski's *Flächeneis*) differ in the extent of their glacierisation and therefore in the degree of control by the underlying relief. Thus continental ice-sheets are almost wholly independent of their foundation while this control in the case of the plateau glaciers and carapaces is most marked. F. E. Matthes[11] distinguished between ice-caps (which include ice-sheets and plateau glaciers) and ice-streams (which embrace mountain, valley and intermontane glaciers).

Continental ice-sheets. For this type there is almost universal assent: it constitutes O. Nordenskiöld's "continental glaciers", Hobbs' "ice-cap type", and the "inland ice" of Drygalski, Ferrar and Gourdon. Early writers, following H. Rink (1851), named it *Isblink* from the peculiar light seen over the inland ice when approached from the sea (see p. 192). The term inland ice has generally been employed in a quantitative sense but has occasionally been used qualitatively, size then being deemed of secondary importance.[12]

The continental "sea of ice" or "ice-inundation"[13] is largely independent of topographical irregularities which are small compared with its thickness. Its low, flat, shield-like dome rests at the margin on high plateaux, as in Greenland, spills over the edge in wall-side glaciers,[14] or drains off as well-defined valley glaciers. To the present representatives in Greenland and the Antarctic, which together form at least 97% of the area of all existing glaciers (cf. p. 94), the Pleistocene added others which buried North America and North Europe.

Greenland. Although Greenland's vast size was suspected as early as the beginning of the 13th century when the *Konungsaspegeln* or *Konungsskuggsjá* (King's Mirror) was written, definite knowledge of it is much more recent, dating from Rink's book of 1857.[15] Dalager's early venture on to the ice in 1751[16] was followed after a long interval by many ascents of the margin, as by A. E. Nordenskiöld[17] in 1882 and 1883 who penetrated 100 km, by E. Whymper[18] in 1867, by Jensen and Kornerup[19] who in 1879 ascended to 1570 m and a distance of 75 km, and in 1893 by Garde.[20]

Nansen[21] was the first (in 1888) to cross the ice from coast to coast (Lat. 64·5°). This feat was afterwards repeated by Peary[22] in 1886, 1892 and 1895, by E. Mikkelsen[23] in 1910, by K. Rasmussen[24] in 1912 and 1917, by A. de Quervain[25] in 1912, by J. P. Koch[26] over the widest part in 1913, by L. Koch[27] in 1921, and by others, e.g. H. G. Watkins[28] (1930, 1931) and A. Höygaard and M. Mehren[29] (1931), at later dates. The routes of transection up to 1930 are shown in fig. 16.[30] An account of these inland journeys was given by L. Koch,[31] especially those in the far north, and a

history of the whole country appeared in the bicentenary volume, *Greenland*[32] (1928). Hobbs has mapped the earlier aeroplane routes.[33]

The ice-sheet (Eskimo: *Sermerssuak*, "big ice") is bordered by the *Yderland* ("outer land") of Rink and the Danes, a strip free from ice, except for small, local masses, and the home of man, musk ox, hare, fox, wild flowers and grass. It consists of both high and low lands and culminates in *c.* 3720 m in the Watkins Mountains in the south-east.[34] Its width on the west has a maximum of 180 km in 62° 30′ N. Lat. and on the east ranges up to 300 km near Scoresby Sound.[35] The strip is narrowest in the south-east and in the north-west. Its total area[36] is 313,000 sq. km, distributed as follows: west, 116,000 sq. km; north, 99,000 sq. km; east, 98,000 sq. km.

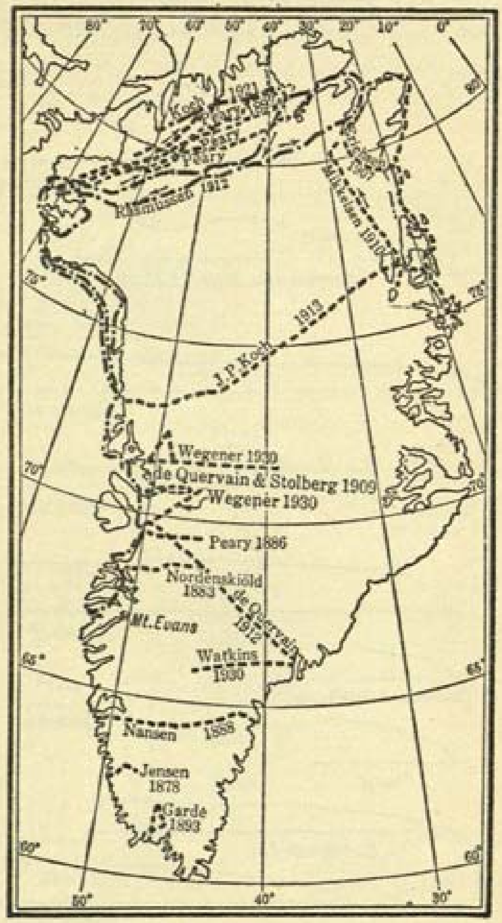

FIG. 16.—Map of Greenland showing routes of transection up to the year 1930. W. H. Hobbs, 769 (2), p. 271, fig. 299.

Of the total area of the country[37] which is about 2·2 million sq. km, 86% or 1,869,000 sq. km is ice[38] (this figure is probably too big) and *c.* 1,650,000 sq. km (earlier estimates were slightly larger[39]) or 75% (637,000 sq. miles) pertains to the ice-sheet.[40] This builds a low, flat dome of regular form (see p. 42) and great thickness (see p. 39), its centre a vast snow plain of magnificent but monotonous desolation, devoid of nunataks, and imperfectly explored. Over much of the interior, the inclination is so slight that it is not detectable by the unaided eye. It varies between 50′ and 2° 4′, but at the margin the surface falls steeply and at an ever-increasing angle as Rink and early explorers noticed.[41] Nansen,[42] for example, found the gradient in feet per mile between sea-level and 1000 m to be, on the east 222, on the west 127; between 1000 and 2000 m 93 and 63; and above 2000 m, 37 and 26 respectively. The calculated angle of slope is as follows[43]:

Between	0–1500 m:	mean 2° 1′ 0″	max. 4° 3′	min. 0° 14′ 0″
,,	1500–2000 m:	mean 0° 20′ 30″	max. 1° 21′	min. 0° 4′ 35″
,,	2000–2500 m:	mean 0° 16′ 0″	max. 1° 1′	min. 0° 4′ 30″
,,	2500–3000 m:	mean 0° 9′ 0″	max. 0° 11′ 20″	min. 0° 4′ 40″

The shape is brought out by the profiles along the routes of transection[44] (fig. 17) which all expeditions, following Nansen's lead,[45] have constructed. The mean height in the west is 2040 m, in the east 2240 m, in the north 1700 m, in the south 1980 m and at the centre 2470 m, giving a mean eleva-

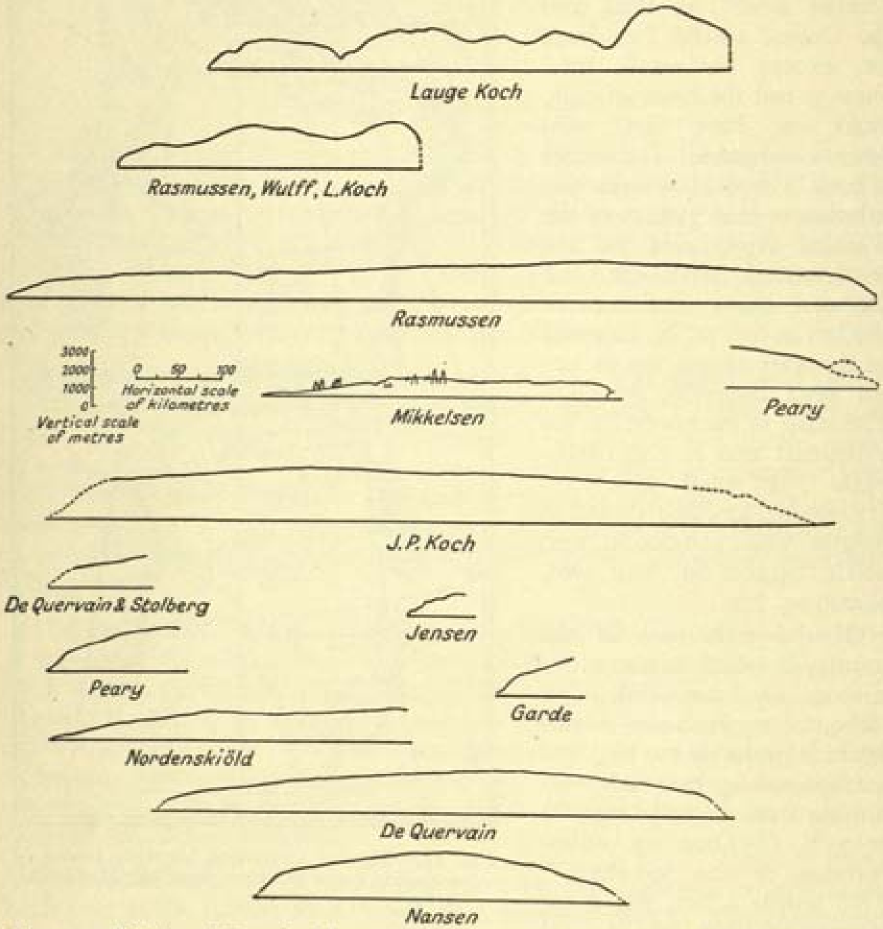

FIG. 17.—Profiles of Greenland ice-sheet along transection routes. W. H. Hobbs, *770*, p. 113.

tion for the whole ice-sheet of 2110 m.[46] These sections have been supplemented by radio soundings from aeroplanes.[47]

O. Baschin[48] expressed the view in 1913 that the ice-sheet had several centres, since the highest points on the various crossings differed appreciably and the divide on J. P. Koch's route lay west and not east of the axis as in the other cases. This suspicion has since been confirmed. Three such domes are now known[49] (fig. 18), a northernmost one, the Central Greenland Dome, at least 3400 m high and situated in 71–77° N. Lat. and 36–37° W. Long. and known from the expeditions of Peary (1892, 1893), Mikkelsen (1910), Rasmussen (1912, 1917), J. P. Koch (1913) and A. Höygaard and M. Mehren (1931); a second one, the Forel Dome, discovered by H. G. Watkins (1931) and con-

firmed by M. Lindsay (1934), which reaches nearly 3550 m at 67° 30′ N. Lat. and 37° W. Long.; and a southernmost and somewhat smaller South Greenland Dome in 65° N. Lat., proved by A. Jessen (1878), O. Nordenskiöld (1883), Peary (1886), Nansen (1888), Garde (1888) and de Quervain (1912). These domes may have been due to differential thinning over buried highlands[50] or to the highlands from which the ice originally expanded[51] (cf. p. 40).

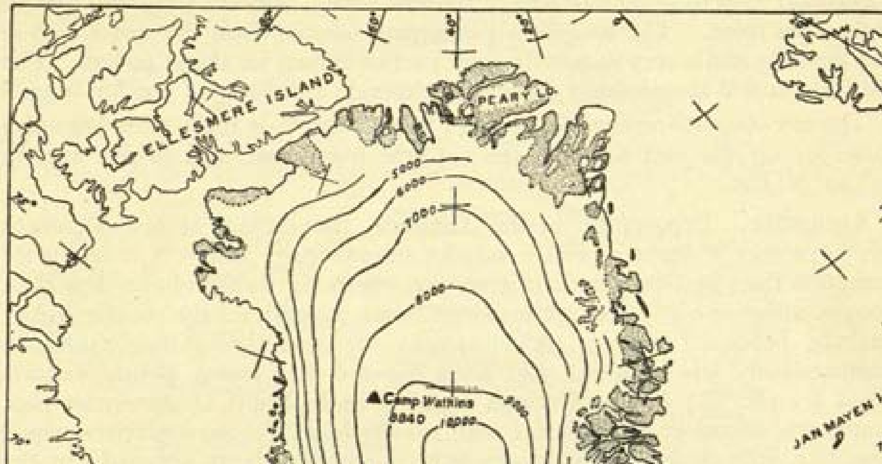

FIG. 18.—Map of Greenland showing the extent and contour (in feet) of the ice-sheet. F. E. Matthes and A. D. Belmont, *T. Am. Gphys. Un.* 31, 1950, p. 175, fig. 1.

The highest points on the several trans-Greenland routes[52] were as follows: Peary, 2225 m; Koch and Wegener, 2928 m; Nansen, 2716 m; Rasmussen, 2225 m; de Quervain, 2510 m; British Arctic Air Expedition, 2804 m. The highest point of the ice-sheet lies north of Mount Forel[53] (3400 m), where proximity to this mountain and high snowfall are favourable.

A depressed fault-zone crossing the country from coast to coast north of

the southern centre may be indicated by the contours of the ice and the topography and geology of the eastern and western coasts,[54] though the available evidence may not fully warrant this conclusion.[55]

Innumerable nunataks dot the ice-sheet towards its margin: they become bigger and more numerous as the edge is approached and ultimately merge into the *Yderland*. Great tongues of ice descend into the fjords. The largest, also the largest in the northern hemisphere, is the Storström in north-east Greenland which has a length of 130 km and the second largest, the Humboldt Glacier,[56] is more than 100 km broad, 100 km long and 100 m high at its front. The longest is Petermann Glacier[57] on the east which is 200 km long and is very thin in its outer part and afloat for about 40 km. The De Geer and Waltershausen glaciers on this side are also over 100 km long.[58]

The *mer-de-glace* reaches the sea on a broad front[59] on three stretches only: these are on the east coast, in the extreme north-east, and in the west in 74° 30′ N. Lat.

Antarctic. Exploration in the Antarctic, the subject of several recent bibliographies,[60] has revealed a vast ice-inundation[61] which is much more complete than in Greenland: it is so overwhelming that probably less than 100 sq. miles or 0·2% are free of ice and even at sea-level the rock is all but entirely hidden. It incompletely masks the underlying topography in comparatively few localities, e.g. King Edward VII Land, South Victoria Land (*c.* 160° E.) and in Graham Land (Palmer Land of American geographers), a land of alpine relief with mountain and plateau glaciers which rises to *c.* 3000 m and is now known to be two major islands separated by the winding Chane Channel.[62] At the margin, the ice gives rise to shelf-ice or floating ice-tongues, or flows freely into the sea as a vast undulating or terraced sheet, normally heavily and deeply crevassed. Sometimes the outlet glaciers form piedmonts or less commonly leave a strip of ice-free ground between themselves and the sea. Most of the coast-line, however, consists of high cliffs (up to *c.* 50 m in height), either "barrier cliffs" or "glacier-cliffs". Marginal nunataks peer through the ice-shroud in numerous places, as in Coats Land (20° W.) and Kaiser Wilhelm II Land (*c.* 90° E.)—off Adélie Land (*c.* 140° E.) and Coats Land the ice-cliff lies near the edge of the continental shelf—and others protrude within the ice-covered plateau, e.g. Neu-Schwabenland[63] (*c.* 3500 m), Dronning Maud Mountains (4500 m), Queen Alexandra Mountains (4600 m), Sentinel Range and Eternity Range in West Antarctica (*c.* 3600 m), Rockefeller Plateau and Edsel Ford Range east of Ross Sea, and Marie Byrd Land east of this again[64] (pl. VA, facing p. 112).

More than 2 million sq. miles (*c.* 5 million sq. km) of the Antarctic continent are still (1953) quite unknown. That the basis is not a frozen sea[65] or congeries of islands[66] (as some thought before the extensive exploration of the 20th century) but, as Captain Cook suspected,[67] a huge continent, is proved by the continuous elevation of the ice; by the nature of the subjacent rocks, as revealed by the erratics, moraines and floor deposits of the Southern Ocean; and by the impenetrable pack-ice along most of the coast. Seiches of three days' duration in Ross Sea hint, it is said,[68] at a channel between the two deep penetrations of Ross Sea and Weddell Sea, dividing the Antarctic into two separate lands, West Antarctica facing the Pacific Ocean between 70° and 75° S. Lat. and East Antarctica (occupying three-quarters of the whole), situated between 65° and 70° S. Lat., Graham Land being the northern tip

of the one and South Victoria Land of the other. It is, however, certain, as C. Markham, Drygalski, Penck and O. Nordenskiöld believed, that a land mass separates these seas.[69] This is suggested by their southerly shoaling—the two seas are shelf seas and not deep gulfs—, by certain distinctive features in their marine faunas, by the wind systems and by geographical investigations which promise finally to eliminate the recurring speculations regarding a possible connexion.[70] Antarctica is a vast elevated massif bordered by a folded mountain system, the "Antarctandes" of Arctowski, which is comparable with the Andes and connected with it through the "Southern

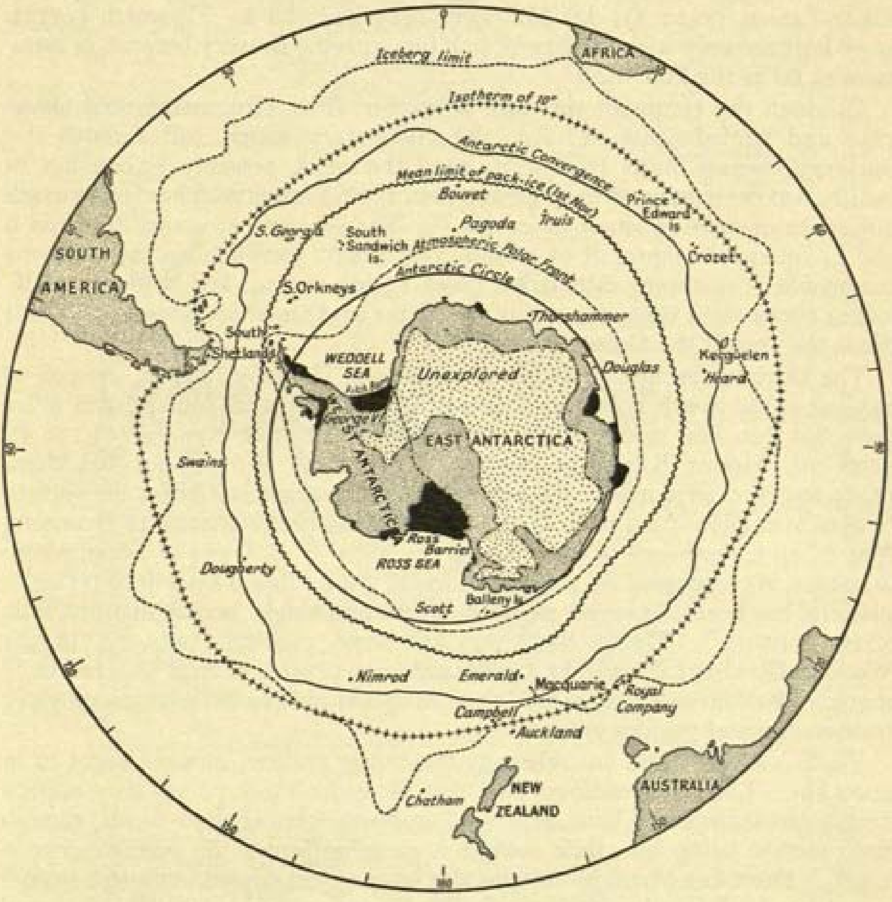

FIG. 19.—Map of the Antarctic. H. P. Kosack, *940*, p. 204 (reduced).

Antilles" of Suess. East Antarctica forms a flat, domed plateau of ice which falls gently northwards from its highest point north of the pole towards the Indian Ocean and steeply towards Ross Sea and Ross Barrier over the 1500 km stretch between Cape Adare (71° S.) and Beardmore Glacier (84° S.). The iceshed, more than 1000 miles (1600 km) long, seems to lie behind South Victoria Land and Dronning Maud Mountains.

While the earlier discoveries,[71] particularly in the past century, were concerned with fixing the position of its coast and emergent lands, some of the

expeditions of the present century—the Belgian (*Belgica*, 1898–1900), British (*Southern Cross*, 1899–1900; *Discovery*, 1901–4; *Terra Nova*, 1907–9; *Antarctic*, 1910–13; *Endurance*, 1914–16; *Quest*, 1921–2; *Discovery* Committee 1923–31), French (*Français*, 1903–5; *Pourquoi Pas?* 1908–10), German (*Gauss*, 1901–3; *Deutschland*, 1910–12), Swedish (1901–3), Norwegian (*Fram*, 1910–12; *Thorshavn*, 1936–7), Australian (*Aurora*, 1911–14), Scottish (*Scotia*, 1902–4), British-Australian-New Zealand (*Discovery*, 1929–30, 1930–1), and the more recent ones which were equipped with reconnaissance aeroplanes, namely H. Wilkins (1928, 1929, 1930), *Norwegia* (1927–8, 1929–30, 1930–1), *Thorshavn* (1933–4, 1936–7), R. E. Byrd (1928–30, 1933–5, 1939–41, 1946–7), Riiser-Larsen (1929–31, D. Mawson (1929–30) and L. Ellsworth (1935), have dealt not only with the coastal belt but with the country behind, in some cases as far as the Pole itself.

Although the continent was first sighted in 1820, circumnavigated about 1840 and landed upon in 1895, the uncertainty which still shrouds the continent's exact limits (as late as 1928 the coast, possibly 15,000 km in length, had been mapped through less than 150° of longitude) has led to much difference of opinion about its extent.[72] Though estimates vary between 9 and 14 sq. km[73] a figure of 13·5 million sq. km (*c.* 5 million sq. miles) seems reasonable[74]—a recent calculation gives 13,101,154 sq. km (without shelf-ice) or about eight times the size of the Greenland ice-sheet or one and a half times the area of the United States.

The altitude averages probably 2250 ± 250 m.[75] It is 2796 m, 2765 m or 2454 m at the South Pole[76] and rises to 2864 m between 100 and 200 miles (160 and 320 km) north of the Pole—Mount Fridtjof Nansen (13,150 ft, 4008 m), Mount Kilpatrick (14,600 ft, 4450 m) and Mount Markham (*c.* 15,200 ft, *c.* 4630 m) are the highest recorded summits. That the surface rises to over 4500–5000 m in Neu-Schwabenland in the interior of Dronning Maud Land, south-east of the Weddell Sea in 81° S. Lat. and 0° Long. where nunataks, accompanied by moraines, divide the ice into a system of reticular glaciers, has been disproved, since the maximum height hereabouts was little above 2710 m.[77] The ice has a very low slope, probably only 2% in the Weddell Quadrant,[78] and like Greenland may possess several ice-centres,[79] though whether and if so where they exist has still to be ascertained since transcontinental profiles are wanting.

South Victoria Land has relatively few outlet glaciers, namely about 12 in 2000 km. Like their analogues in west Greenland (see p. 39) they occupy deeply cut valleys and have depressed amphitheatres at their heads, though their motion being less, their surface is more uniform—the average slope is 1–4%. Dronning Maud Mountains also have a great array of outlet glaciers,[80] e.g. Liv, Axel Hamberg, Kent, Isaiah Bowman, Amundsen, Thorne and Leverett glaciers, as have the Edsel Ford Mountains.[81]

Highland ice-caps. Highland ice-caps are miniature replicas of ice-sheets but differ from these in size and method of alimentation. Like island ice-caps, they may mark stages in the dissolution of larger masses. Peary Land in north Greenland and East Friesland, Spitsbergen, may be taken as examples, though the latter, O. Nordenskiöld's original "Spitsbergen type",[82] is transitional between the Highland ice-cap and plateau glacier.

Island ice-caps. A highland ice-cap perched up on an island may be termed an island ice-cap; it is Nordenskiöld's "ice-cap" and Ferrar's "local

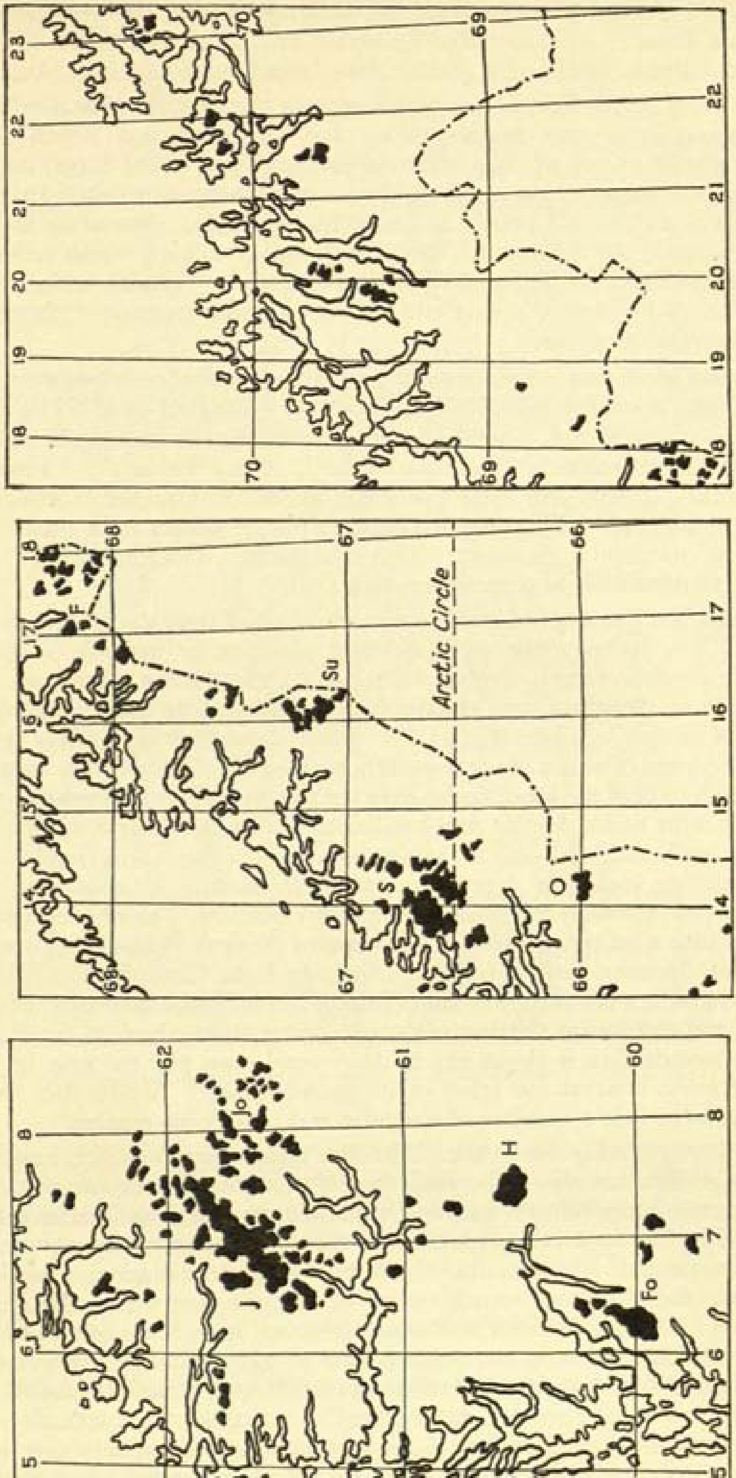

FIG. 20.—Distribution of the snowfields and plateau glaciers of Norway at the beginning of this century.

ice-cap". Examples are supplied by the larger islands off west Greenland, by Baffin Land,[83] by Severnaya Zemlya[84] and North-East Land,[85] by the islands of Franz Josef Land, and by Ross Island and Snow Hill, Antarctica.

The ice of North-East Land, which crowns plateaux about 400 m high and is divisible into three masses, West Ice, East Ice and South Ice, has a total area of 11,425 sq. km. East Ice is the largest of the three, its surface in the centre rising to 730 m where the ice may be 250 m thick. On the east and south it thins off evenly and is extremely thin. Nunataks and other features prove the thinness of West Ice (possibly 125 m) which reflects to a considerable degree the underlying topography. Vertical terminal cliffs, up to 50 m high on the east and south, suggest extensive glacierisation and Antarctic conditions.

Plateau glaciers. A plateau glacier reposes on a more or less even surface but, having a smaller alimentation, is not so embossed as the Highland or Island ice-caps; it may, indeed, have no embossment or may sag in the centre. It constitutes Richter's[86] "plateau type", P. C. Visser's[87] "Firnplateau type", the "Norwegian type" of some authors[88] and the "Scandinavian type" of others.[89] Like the preceding types, it differs only quantitatively from the continental ice-sheet. This relationship, though denied,[90] has led Hess[91] to repudiate its separate existence.

Plateau glaciers depend much more upon relief than do ice-sheets or ice-caps. They sprawl over small elevated plateaux or uplands which have scarcely any intervening divides. They have, therefore, a central or common névé (Norw. *Sneefond*) and usually lack surface-debris since nunataks, save near the margin, are rare (fig. 21).[92] Their domed surface differs strikingly from the form of valley glaciers as early noticed.[93] Their flow is outwards to the margin where the ice cascades over the rim in narrow tongues (Norwegian glaciers with suffix *bre* or *brae*) radiating down the valleys (Jostedalsbrae has 24 such tongues) and in some cases, as on the Vatnajökull,[94] having "intakes" or ladle-like depressions at their heads. In this way, plateau glaciers pass through transitions into valley glaciers. Examples, needlessly erected into a subtype,[95] are the Oxtinder of Norway,[96] and the Ebensferner (west of Brenner) and Glacier of Mont de Lans (Dauphiné) of the Alps. Plateau glaciers are relatively thin: Jostedalsbrae[97] which consists of a series of isolated firn basins flowing outwards is not more than 20–40 m thick—Tungsbergsdalsbre is about 320 m deep—and even the thickest in Iceland cannot quite conceal the relief of its foundations.[98] By further thinning, they dissolve into a number of irregular and patchy ice-masses.

Norway probably best exemplifies the type, though valley, hanging and cirque-glaciers are also to be seen there[99]: cirque-glaciers, for instance, are prevalent in Jotunheim[100] and on Dovrefjeld, as well as in Sweden (cf. Hamberg's map[101]) and valley glaciers in Sarek—Hamberg (1910), mentioned *c.* 220 corrie and valley glaciers. Nevertheless, plateau glaciers, probably only thin, are the chief type north of 60° N. Lat. (fig. 20) where the fjords, by drawing in the air, concentrate the precipitation upon the summit uplands.[102] They rest upon the palaeic surface (see p. 348) and include Svartisen,[103] Jotunheim,[104] Folgefonn,[105] Sukkertoppen[106] and Jostedalsbrae.[107] Their areas in square kilometres are as follows[108]: Svartisen, 450 (with the adjacent masses, 732·8); Jotunheim, 207; Folgefonn, 288 (alternatively 220, 264, 290 or 300); Jostedalsbrae, 1076 (*c.* 10–12 km broad and 90 km long), the biggest

glacier on the European mainland; Jökelfjeld (the northernmost plateau glacier: 70° 10′ N. Lat.), 190; and Sulitelma, 52.

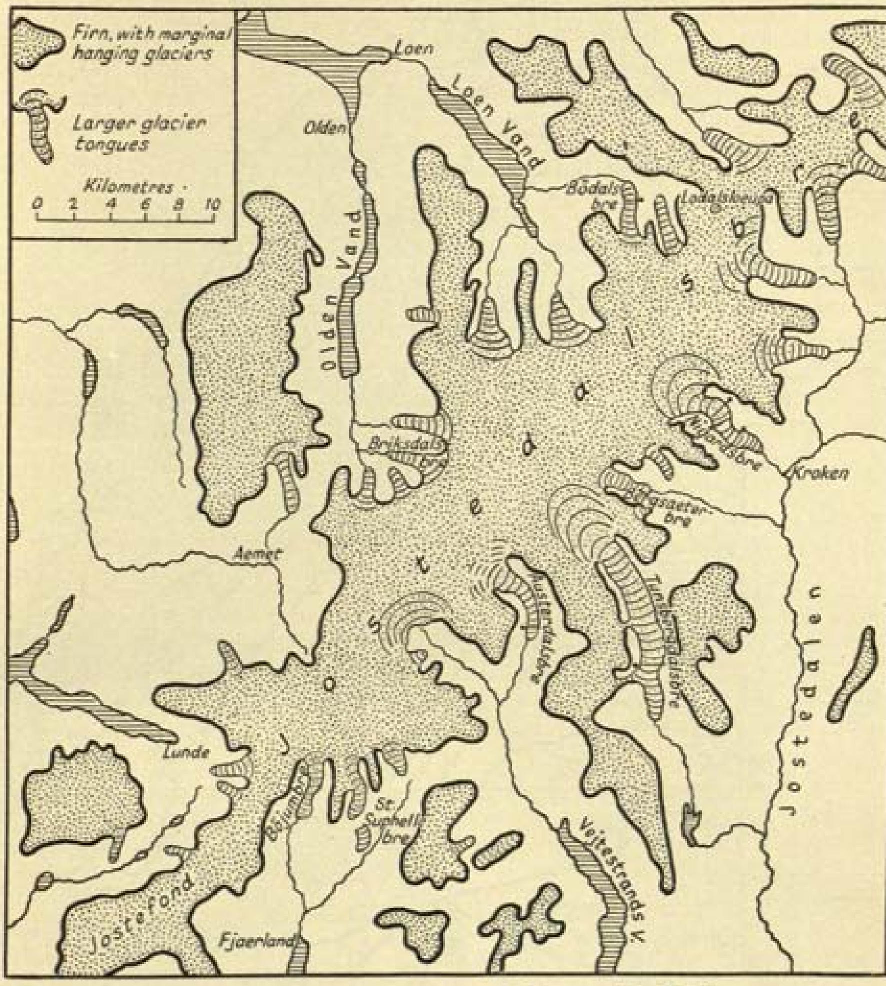

FIG. 21.—Map of the Jostedalsbrae, Norway. R. Finsterwalder, *J. Gl.* 1, 1951, p. 557, fig. 1.

Plateau glaciers also exist in Iceland[109] (fig. 22). These *jöklar* (sing. *jökull*)—this Icelandic word applies indiscrimately to snowfields, glacier tongues and isolated ice-covered mountains—are relatively large as the following figures[110] (in square kilometres) indicate: Vatnajökull, 8410; Hofsjökull, 1350; Langjökull, 1300; Mýrdalsjökull, 1000; Drangajökull, 350 (160 ?). Vatnajökull[111] (fig. 23), the largest glacier in Europe and in extra polar regions and more than the equivalent of all other Icelandic glaciers, has a southern edge which is unrestricted and a northern one which abuts against rising ground. While Thoroddsen's estimate for Vatnajökull may have to be raised to 8800 or 9000 sq. km[112] his figure for Hofsjökull should, it is said, be reduced to 700[113] and that of Hofsjökull, Langjökull and Túngnafellsjökull be diminished by 1000.[114]

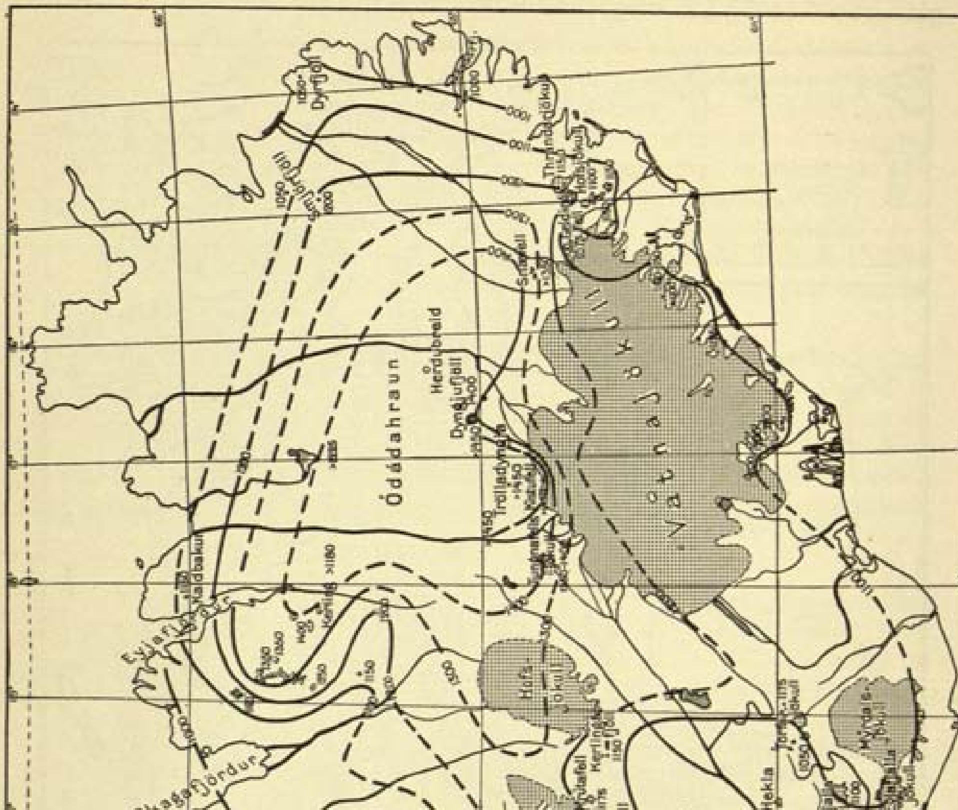

FIG. 22.—Map of Icelandic glaciers and snowline. H. W : son Åhlmann, *21*, pl. iv, after p. 208.

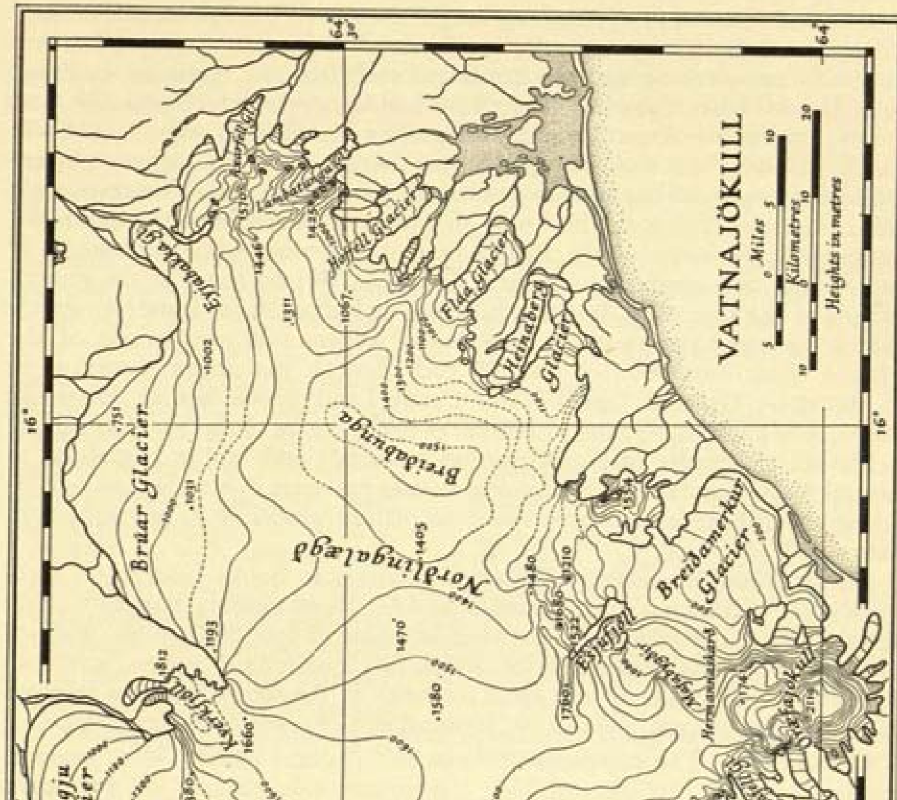

FIG. 23.—Map of Vatnajökull. H. W: son Åhlmann, *25*, p. 11, fig. 11.

The plateau type is represented too in the Übergossene Alm, Marmolata Glacier and Claridenfirn (area 8 sq. km) of the Alps[115]; in Barents Island and Edge Island, Spitsbergen; in the north island of Novaya Zemlya[116] where the ice is 400 km long and 100 km broad and rises to *c.* 1000 m; in Franz Josef Land,[117] that region in the northern hemisphere which climatically most nearly approaches Antarctica (the temperature in July averages 1·3° C, in January, −26·5° C, and during the year −14·1°C)[118] and where glaciers which cover 91% of the surface and rise to 300 m above sea-level frequently end in high sea-cliffs and where only 36 species of flowering plants are known to grow[119]; on some of the islands of east and west Greenland south of 61° N. Lat. and commonly in north Greenland[120]; on some of the eastern islands of the Canadian Arctic archipelago including Baffin Land[121]—gravimetric surveys have proved a thickness of 467 m and calculations of the inward shear-force exerted by the rock floor a maximum thickness of *c.* 460 m—, Grinnell Land[122] and Grant Land which has an ice-shield 800 sq. km in extent[123]; and in Alaska, as in the centre of the St. Elias Range.

Plateau ice occurs in Nan-Shan (40 km long) and in the Himalayas in central Asia,[124] and a form transitional to valley glaciers in central Caucasus[125] where the Elbruz has apparently a true plateau glacier[126] *c.* 200 sq. km in extent.

The southern hemisphere has representatives in South Georgia,[127] Kerguelen[128] (Richthofen Ice), Heard Island,[129] Royal Society Range[130] (Blue Glacier), and in Patagonia[131] between latitudes 46° 30′ and 47° 30′ and 48° and 50° 30′. The Patagonian ice, which fills longitudinal valleys, stretches for 700 km and breaks up into small ice-fields in the south, has been referred to a "Patagonian type"[132] and by Nordenskiöld[133] to his Spitsbergen type.

Carapaces. The carapaces of Tyrrell[134] ("dome glacier"[135]; "tropical glacier type"[136]; "glacier-cap"[137]; "mountain side glacier"[138]) surmount the rounded or flat tops of isolated mountains which are little dissected. Their ice, virtually moraineless, is divergent, though insolation, slope and conditions of snowfall often make the position eccentric. They crown some of the tropical mountains,[139] such as Ixtaccihuatl (5286 m) and Kilimanjaro (6160 m), and a few rounded mountains in Spitsbergen[140] and North America[141] (e.g. Mount Jefferson, Mount Hood, Mount Adams and Mount Rainier—this has the longest and largest ice-stream (*c.* 10 km long), viz, the Emmons Glacier, in the continental United States). The ice on Bouvet Island,[142] about 30 sq. miles (*c.* 80 sq. km) in area and 935 m high, should probably be classed among the carapaces.

2. *Mountain Glaciers*

The second main group, Nordenskiöld's "mountain glaciers", are found in mountainous regions and fall naturally into a series of types of decreasing glacierisation.

Reticular glaciers. The first of these, the reticular glaciers of Tyrrell, is Nordenskiöld's "Spitsbergen type", Garwood and Gregory's "confluent series of glaciers"[143] and Penck's *Eisstromnetz*[144]: Drygalski[145] distinguished an *Eisstromnetz* with common reservoir and separate tongues and a *Talnetz* with separate reservoirs and united tongue.

Confluent glaciers of the reticular type fill whole valley systems and exhibit

in their flow a certain independence of the relief. They stream over passes and cols and deploy sometimes as piedmonts. They serve as a transition between valley glaciers and ice-sheets.[146] H. Louis's[147] observation that mountainous areas projecting 100 m above the snowline have such a net in U.S.A. but not in the Alps shows that the controlling factor is the extent of dissection and not the height of the mountains above the snowline.

Reticular glaciers of to-day are well developed in Novaya Zemlya[148] and Spitsbergen[149] (together with valley and piedmont glaciers) and in Neu-Schwabenland, Antarctica.[150]

Dendritic glaciers. The dendritic glaciers of Hobbs,[151] united with the reticular type and named "transection glaciers"[152], are a system occupying a trunk valley and its branches. The type was based upon M. Conway's description of the Baltoro (66 km; fig. 24), Biafo (60 km) and Hispar (59 km) glaciers of the Karakoram Mountains[153] and fits others since found there[154]—the eight longest glaciers in central Asia (Fedchenko, Siachen, Inylchek, Hispar, Biafo, Baltoro, Batura and Koi-Kaf) are the only glaciers outside the subpolar regions which exceed 30 miles (*c.* 48 km) in length. The type occurs too in Alaska while the Tasman Glacier of New Zealand, which is *c.* 18 miles (29 km) long, provides an example from the southern hemisphere.

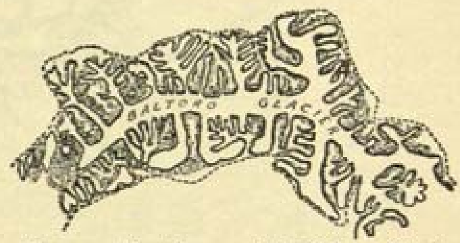

FIG. 24.—Sketch-map of the Baltoro Glacier. W. H. Hobbs, *769* (2), p. 385, fig. 410.

Valley Glaciers. The valley glacier[155] of Richter and Hochstetter (Norw. *Dalbrae*) is the Alpine glacier of many classifications and historically the archtype. The Alps, principally in the west and centre (Mont Blanc Group, Pennine Alps, Bernese Oberland) boast altogether some 2000 glaciers of one type or another,[156] their lengths varying greatly as Heim's and R. v. Klebelsberg's figures show.[157] The areas of the most important (in square kilometres) are: Aletsch, 129 (1880: 115; 1945: 100; length, 25 km), Gorner, 67; Mer de Glace, 55; Fiesch, 41; Lower Aar, 39; Lower Grindelwald, 29·5; Pasterzenkees (the largest glacier in the eastern Alps), 24·5; the rest are nearly all below 20 or even 10 sq. km.

The Fedchenko Glacier in the Pamirs, the longest outside the subpolar region, is 75–77 km in length[158] (area, 1350 sq. km). The longest in the Karakoram Mountains[159] is the Siachen, 75 km long (area, 1553 sq. km) and Inylchek Glacier of Tienshan 71 km long, the Batura Glacier of the Hindu Kush 60 km long,[160] and the Zemu Glacier of the Himalayas 25 km long.[161]

Valley glaciers head in one or more névé fields or in sheet-ice and depend for their shape upon topography and valley windings. They are usually longer than broad, though exceptions are numerous. They may project at their snout, as in some of the hanging valleys of Yakutat Bay region,[162] or bend over a steep declivity, as at the end or along the side wall of a valley, e.g. in the Rhône Glacier: these form Hamberg's "glacier with hanging glacier-end".[163] They may also unite over cols, as in the Ferrar and Taylor glaciers, Antarctica,[164] or build composite glaciers if two or more streams unite, as in the

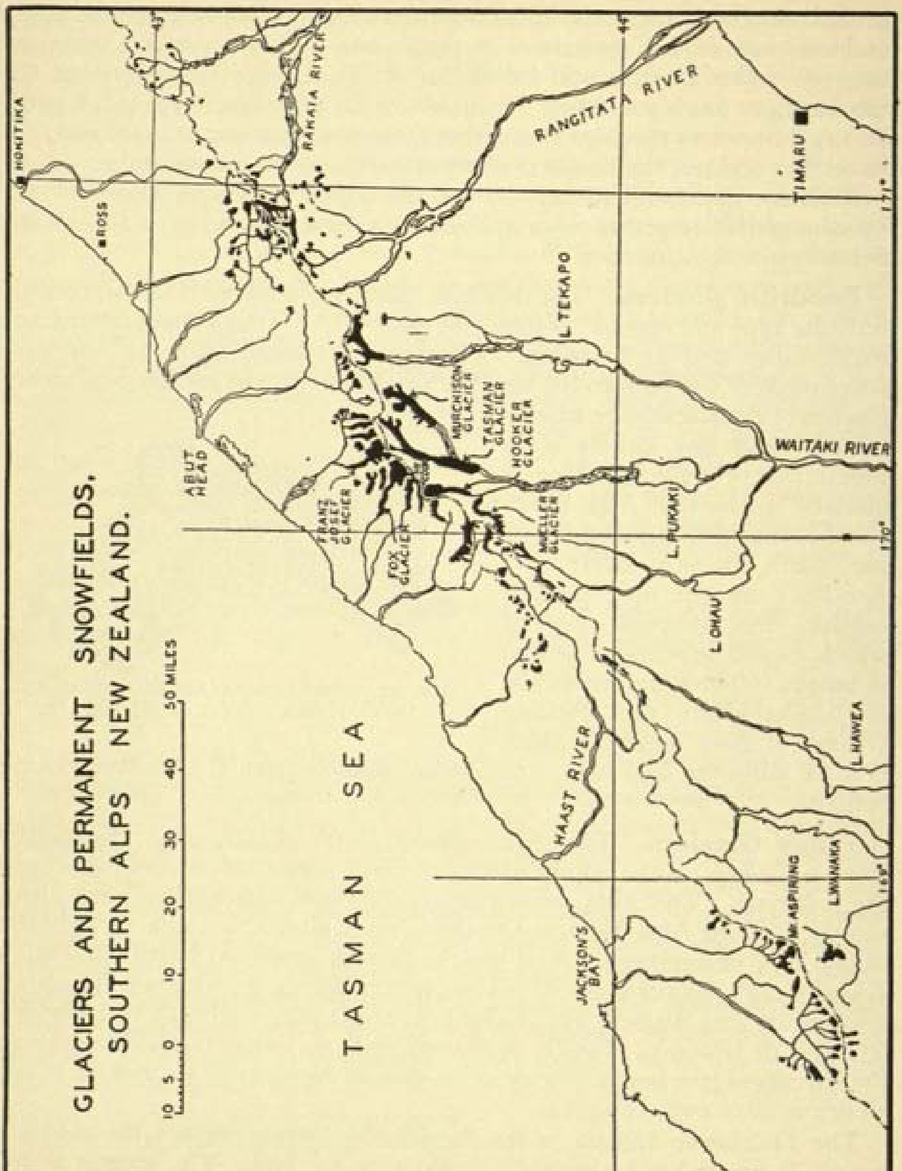

FIG. 25.—Glaciers and permanent snowfields of the Southern Alps of New Zealand. R. P. Suggate, *J. Gl.* 1, 1950, p. 423.

Aletsch, Mer de Glace, Hintereisferner and Vernagtferner of the Alps. These are common in the Caucasus, Himalayas and New Zealand[165] (fig. 25): they constitute de Martonne's "Himalayan type".[166] The glaciers of New Zealand are best developed about Mount Cook (3813 m). To the west, the most important are the Franz Josef and Fox glaciers, to the east the Hermitage, Müller, Tasman, Hooker and Murchison glaciers.

Valley glaciers may be subdivided into several type-groups according to the size of their basins in relation to their tongues.[167] The commonest and

most typical, exemplified by the Rhône, Hintereisferner, Allalin and Forno glaciers, shows an area-distribution curve which rises steadily from a low value to a pronounced maximum whence it falls to a fairly low value again; the second group, e.g. the Great Aletsch Glacier, has a maximum higher up; the third group, e.g. the Siachen Glacier of central Asia and the Styggedal Glacier of Norway, has no proper firn; and a fourth group, commonly represented in Spitsbergen, reaches its maximum area at a low altitude.

A tributary glacier usually flows along side, less commonly upon a major glacier.[168] If the lateral glacier is steep, it may fall on a trunk glacier as a remanié glacier[169] (see p. 91) or push the trunk glacier to the farther side of the valley,[170] ridging it, or even penetrating beneath it.[171] In other cases,[172] where the two rock-floors are markedly discordant, the tributary may flow on

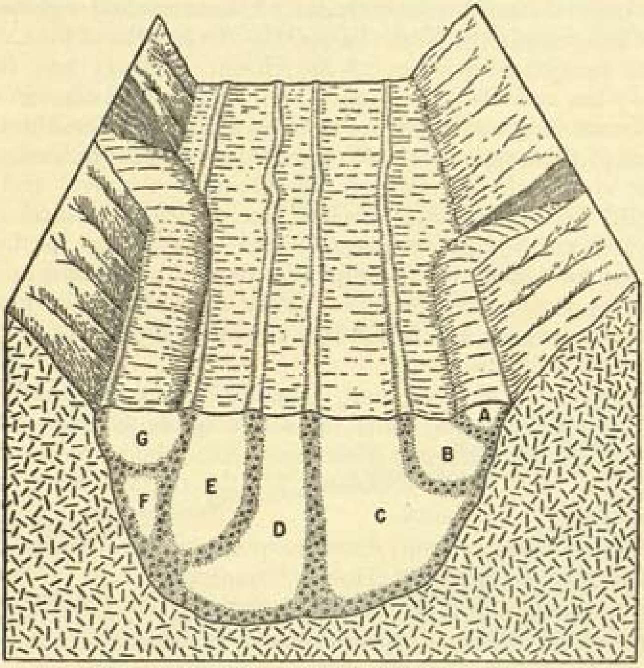

FIG. 26.—Diagram of the parallel, inset and superimposed ice-streams in a composite glacier. R. P. Sharp, *J. Gl.* 1, 1948, p. 184, fig. 4.

to the master glacier as a "parasitic glacier", overriding the surface moraine and shearing it, striating by its own motion the upper surface of the lower ice but generally displaying no such signs of slipping. The parasitic glacier may be torn off by the lower glacier and broken and be ultimately reduced to ice-pyramids by melting,[173] though "cold" polar glaciers may retain a superimposed ice-stream for a much longer time.[174] The confluent glaciers retain their own banding[175] but otherwise move together as a unit, the crevasses penetrating them both, irrespective of the separating plane,[176] the upper glacier moving as the resultant of its own motion and that of the glacier beneath.[177] If the valley floors are initially discordant and the trunk glacier is thick enough to raise a barrier across the tributary valley, an inset relationship develops[178] (fig. 26).

A parasitic glacier deposits its ground-moraine and terminal moraine upon the main glacier,[179] its lower end being marked by a dirt-band or a "crescentic moraine" (also termed "cross", "oblique", "horse-shoe", "melt" or "pseudo-terminal" moraine[180]). Two, three or four such glaciers may rest one upon the other over the whole breadth.[181]

Valley glaciers are the principal type in the temperate zone, as in the Alps, Caucasus,[182] Himalayas[183] (these have more than 1000 glaciers of an area of 1968 sq. km, of which 502 sq. km are on the south side), Tienshan, Andes,[184] Rocky Mountains and Canadian Cordillera.[185] In New Zealand[186] they are restricted to mountains above *c.* 2286 m in height in the Southern Alps between the vicinity of Milford Sound (44° 47′ S.) and Arthur's Pass (42° 53′ S.), though permanent or nearly permanent snowfields occur for a further 145 km to the north-east in Spenser Mountains and a glacier occupies the crater of Ruapehu in the North Island[187]—the lengths of the main glaciers are Tasman 29 km, Murchison 18 km, Franz Josef 15 km, Müller and Godley *c.* 13 km and Hooker 11·7 km.[188] They also occur in the South Island of Novaya Zemlya[189] and are possibly to be numbered in their thousands or tens of thousands on the Wrangel Mountains and Alaska Range.[190] The biggest are in the Karakoram Mountains (see above) and in South Victoria Land where the outlet glaciers (see p. 78) in general increase in size southwards and the Beardmore Glacier,[191] the largest in the world, is 120 miles (194 km) long, 15–35 miles (24–57 km) broad and at least 5000 sq. miles (*c.* 13,000 sq. km) in area. Other Antarctic outlet glaciers have also considerable lengths, e.g. Amundsen 160 km, Thorne 145 km, Drygalski 120 km and Ferrar 80 km.

In high latitudes, as in Greenland, Alaska,[192] Ellesmere Land[193] and Jan Mayen,[194] valley glaciers may enter the sea or fjords as "tide-water"[195] or "tidal"[196] glaciers, some like the Malaspina Glacier being piedmont glaciers, others dendritic. In "subtidal glaciers"[197] the edges rest on land over a considerable width of the valley.

Where glaciers protrude from inner ice-plateaux through mountain gaps, they form "through glaciers"[198] (Hobbs' "transection type"); two dendritic glaciers join to produce Hobbs' "twinned glacier type".[199]

Cirque-glaciers. Cirque-glaciers (Norw. *Botnbrae*), which occupy cirques, have been called the "Pyrenees type" since they especially distinguish that range,[200] particularly its northern slopes between the Garonne and Val d'Assone; only the Vignemal approaches the shape and dimensions of a true valley glacier. Owing to the snowline's great altitude cirque-glaciers are naturally the main type in the tropics.

These glaciers may end within the cirque or thrust their snouts out on to the side or upper end of a lower valley as "spill-over glaciers".[201] On the other hand, they may shrink to the foot of the cirque-wall as "horse-shoe glaciers",[202] i.e. broad glaciers with incurving fronts, as exemplified in the Glacier National Park. Most of the glaciers in Sweden[203] and North America[204] (outside Alaska) are of this kind.

Very rarely a glacier may inherit a volcanic crater: the Cerro alto glacier of Ecuador is an instance of such a "cauldron glacier".[205] Kamchatka[206] has cauldron-glaciers, star-glaciers and other types depending upon volcanic relief.

Hanging glaciers. Hanging glaciers (Norw. *Haengebrae*), Supan's "oro-

graphic glaciers",[207] were one of the first types to be differentiated: they were de Saussure's "glaciers of the second class".[208] Usually small and steep and frequently as broad as long, they are plastered against the walls of precipices, as on the northern face of the Jungfrau, or nestle in hollows in a mountain flank or the escarpment of a plateau. They characterise the Caucasus and grade into "niche glaciers",[209] which lie in niches in steep-sided peaks or ridges (*Jochgletscher*, or "glaciers of the third order" of F. Simony), or pass into "cornice glaciers"[210] or "cliff glaciers"[211] like those which are so numerous among mountain glaciers in western North America.[212] There are transitions too to the "cascade glaciers"[213] which protrude from the mouths of hanging valleys or "wall-side glaciers"[214] (*Stufen-* or *Flankenvereisung*)[215] which depend from portions of highland or other mountain ice and rest on the rock-terraces of valleys and plateaux,[216] as in north-east Greenland and those parts of Spitsbergen which have horizontal sedimentary strata of varying hardness. Such glaciers adhere to the mountain shelves since they are frozen to the rock. Cascade glaciers are also known from Alaska[217] and from Lapland where they repose upon old shore-lines.[218]

The smallest of all glaciers are the "snow-bank glaciers" (fig. 27), the

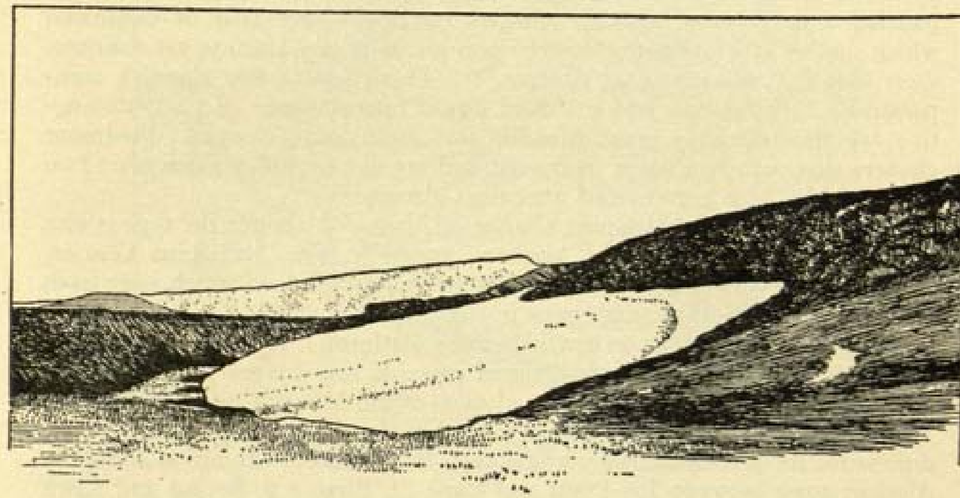

FIG. 27.—Snowbank glacier and nivation hollow, Quadrant Mountain, Yellowstone National Park. W. H. Hobbs, 769 (2), p. 368, fig. 390.

"drift glaciers" of T. C. Chamberlin.[219], the "temporary glaciers" of E. Collomb,[220] "snow-drift ice"[221] or Hobbs' "nivation type". They occur, for example, in east Greenland and line the sea in the Antarctic where they have been named "ice-foot glaciers"[222] by Nordenskjöld, Gourdon and Pirie or "suspended coastal glaciers" by Arctowski. Similar forms are the "wind-glaciers" of Neu-Schwabenland.[223] To this type also belong the majority of the small glaciers in the south Tyrol Dolomites, in the North Limestone Alps and in the Carnic and Julian Alps.[224]

Perennial snow-banks which imitate glaciers in their shape accumulate in the lee of hills, at first rapidly, later more slowly as they approach a certain critical angle.[225] They are independent of the snowline and may be at all elevations down to the low altitudes of the coasts or lie in river beds.[226]

Being inert, they have neither banding nor dirt-bands. Every transition from motionless snow-bank to moving snow-bank glacier occurs in west and north Greenland.[227] The requisite thickness for the transition depends upon pressure, temperature and the gradient of the underlying bed,[228] and most probably exceeds 30 m[229]; Matthes averred that on a slope of 12% it was 38 m (see p. 303).

Snow-banks, permanent or transitory, are innumerable about the world's snowlines. They line hundreds of kilometres of the Antarctic ice-border and build aprons at the end of the outlet glaciers in Ross Quadrant. Snow-bank glaciers also are numerous in the Antarctic, e.g. in South Victoria Land,[230] in Grant Land[231] and Colorado,[232] and at the glacier snouts in west and north Greenland[233] where their length approaches 5 km and their tiny grains reveal their parasitic nature. Europe's most southerly glaciers, namely those in the Sierra Nevada, are of this type[234] (cf. p. 15).

3. *Lowland Glaciers*

Piedmont glaciers. Most important among the lowland glaciers are the piedmont glaciers, first recognised by Russell.[235] These arise by the expansion and union of adjacent glaciers (mountain, reticular or dendritic) which deploy at a mountain foot or upon plains of any altitude yet maintain their identity as commensal streams.[236] They have a low slope, a comparatively sluggish flow and a surface almost uncrevassed. Equally distinctive are the relatively great breadth and multilobate margin. Piedmont glaciers that occupy a basin or trough and are fed by valley glaciers on two or more sides have been named intermont glaciers.[237]

The original is the Malaspina Glacier of Alaska[238] whence the type is also designated "Malaspina" or "Alaskan" type.[239] The Malaspina Glacier, produced by a coalescing of nearly a dozen tributary glaciers which issue from a vast inland icefield through gaps in the coastal St. Elias Range (5955 m), spreads out as a broad fan on to a wide shore platform. It is 305–458 m thick, flows over a surface which is inclined inwards and covers 1035 sq. miles (2682 sq. km), i.e. more than all the glaciers of the Alps put together (fig. 28).

The type is now confined to high latitudes where obviously alone the conditions for its development are available. Its glaciers are common along the Alaskan coast between Icy Point and Cape St. Elias, e.g. Bering and Alsek glaciers.[240] Their broad, symmetrical bulbs expand over the smooth coastal plain to the tidal water or fail to reach it because of morainic tracts covered by dense spruce forests or wide barren stretches of outwash gravel. They are also known from the Canadian Rocky Mountains (e.g. Wenkchemna Glacier[241]), both sides of the Evigheds Fjord of Greenland,[242] Ellesmere Land,[243] Prince Charles Foreland, Spitsbergen,[244] Novaya Zemlya,[245] Franz Josef Land,[246] Severnaya Zemlya,[247] and in the southern hemisphere[248] from New Zealand, Patagonia and South Georgia. On the Antarctic continent, they deploy upon the coastal belt of South Victoria Land[249]—Wilson Piedmont (breadth, 36 miles: 58 km; average depth, 5 miles: 8 km) and Butter Point Piedmont are examples—and make much of the Antarctic edge almost inaccessible.

Near sea-level, these glaciers may pass by excessive precipitation into floating ice-tongues and shelf-ice but by diminished precipitation may grade into expanded-foot glaciers at this or any elevation.

Expanded-foot glaciers. The expanded-foot glacier[250] (=Åhlmann's "foot glacier") emerges from a mountain valley as a fan and debouches upon a wider trunk valley or upon a plain which may be an elevated plateau or a raised marine platform. By no means common, except in Alaska,[251] it is known from Sverdrup Land,[252] north Greenland,[253] Spitsbergen[254] and Antarctica, e.g. South Victoria Land,[255] where the snowfall is heavy and the snowline high. The Rhône Glacier had an expanded foot before its retreat after 1870.

Reconstructed glaciers. Glaciers retain their continuity on considerable slopes: in the Alps,[256] the Glacier des Bossons has one of 28°, the Glacier de Taconnaz 29°, the Eiger Glacier 34°, and the Giessen Glacier 39°. If, however, the slope becomes steeper the glacier breaks off in ice-avalanches (see p. 22).

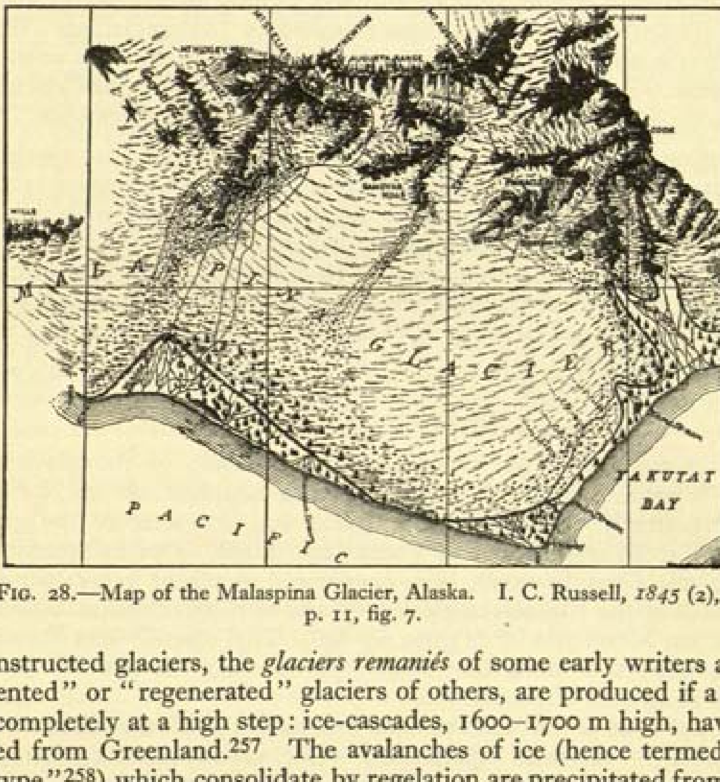

FIG. 28.—Map of the Malaspina Glacier, Alaska. I. C. Russell, *1845* (2), p. 11, fig. 7.

Reconstructed glaciers, the *glaciers remaniés* of some early writers and the "recemented" or "regenerated" glaciers of others, are produced if a glacier breaks completely at a high step: ice-cascades, 1600–1700 m high, have been described from Greenland.[257] The avalanches of ice (hence termed "avalanche type"[258]) which consolidate by regelation are precipitated from either cirque or valley glaciers or from sheet-ice of elevated lands. Exceptionally, the reconstructed glaciers may form by wind drift, as in north Greenland, or by avalanched snow or névé and in turn may nourish other glaciers of the same type. "Ice aprons" may unite them with the ice above.[259]

Reconstructed glaciers are rare since ice-avalanches usually fall into the sea, as in west Greenland, into a lake, or into warm regions where they soon melt. Those that do occur are merely a few hundred metres long and a few metres thick. Nevertheless, they have all the characteristics of a true glacier, including movement and a stratification with dirt layers. They are found if the country is steep and high, as along the coastal cliffs[260] of Greenland or

Norway where plateau glaciers are prone to this development. A well-known example is Suphellebrae[261] in the Sognefjord which is 100–200 m thick and 1 km long and is fed by ice-avalanches from Jostedalsbrae over a fall of 200 m. Others occur on Lyngenfjord and below Jøkelfjeld and Svartisen: they include the Kjendals–Briksdalsbrae and Bojumbrae of Jostedalsbrae. In the Alps, where they are less numerous than they were before the retreat of the last 100 years, the best known line the north wall of the Jungfrau massif: the Lower Schwarzwald Glacier of the Wetterhorn (Bernese Oberland) is an instance.

The "fan glaciers" of Yakutat Bay region[262] which are related forms are nourished from snows which glide down steep and narrow chute-like valleys to the more or less flat ground below.

Ice-slabs. Ferrar's ice-slabs,[263] discovered by him in South Victoria Land between Koettlitz Glacier and the vicinity of Drygalski Ice-barrier tongue (32 km), are one of the forms assumed by "dying glaciers".[264] Their large inert masses ("glacier-outliers")—the Blue Glacier is 20 miles (32 km) long—arise from the dead aprons of piedmont glaciers or from glaciers which because their supplies have failed have parted from their firnfields.

Fringing glaciers. In certain places in the Antarctic, for example along the west coast of Graham Land, there is a belt of low-lying glaciers truncated on the outer side, at the boundary between the sea and rock, by ice-cliffs which are remarkably constant in height. These fringing glaciers[265] (= Gourdon's "piedmont glaciers", Holtedahl's "strandflat glacier") are apparently relics from the time when shelf-ice filled the bays off the coast.

Forms transitional between land-ice and sea-ice are described at a later page (see ch. VII).

4. *Geographical Distribution*

Our knowledge of the geographical distribution of the glaciers of the world,[266] great though it is, is still far from complete: we are, for example, only imperfectly informed of the extent and character of the glaciers of Canada's arctic islands and of Franz Josef Land. Our information on the Asian glaciers has, however, been augmented in recent years: surveys have been made of the Himalayan glaciers[267] and of those of central Kurdistan,[268] Karakoram Mountains[269] (13,660 sq. km), Tien-shan[270] and Russian Altai (230 sq. km) and Mongolian Altai (170 sq. km), the Altai[271] having 407 glaciers of an area of 590 sq. km.

Ten per cent of the land-surface or about 3% of the earth's surface is glacierised[272]—the term "glacierised" (Ger. *vereist*) has been introduced[273] for land covered with ice as distinct from land glaciated by ice. The Greenland and Antarctic ice-sheets together form 97% of the area, 99% of the volume of the world's land-ice.

As the snowline falls progressively from low to high latitudes, its curved plane intersects lower and lower lands. Hence the types encountered in passing from the tropics to the poles are generally those of increasing glacierisation; cirque, hanging and small valley glaciers and the carapaces of the Andes and volcanic mountains of tropical Africa (where high altitude compensates for high latitude) give place to piedmont glaciers in Alaska and to plateau glaciers in Norway, Iceland, south Greenland, Kerguelen and Heard Island. These in turn yield in yet higher latitudes to highland ice in Spits-

bergen and west Greenland and to continental ice-sheets in Greenland and the Antarctic—the general absence of ice from the Canadian Archipelago and north Siberia may be due to the low precipitation (see p. 643) or to low relief.[274]

The mountains of Europe may be subdivided glacially into three categories[275]: those which lie definitely above the snowline and harbour snowfields and glaciers (Iceland, Norway, Alps, Pyrenees); those which fall short of the snowline by a narrow margin and have permanent snowfields or incipient glaciers (Grampians, Sierra Nevadas, northern Urals, Carpathians): and those which, though definitely below the snowline, retain snow-drifts at particular spots to a late date in summer (Abruzzi Apennines, Mount Etna, Balkans and Snowdon).

The change with latitude is most conspicuous in North America.[276] North of the southernmost glacier (Palisade Glacier) of the U.S.A. in the Sierra Nevada in 37° 6′ N. Lat.,[277] there are valley and cirque glaciers which advance but little below the snowline. In north California, Oregon and Washington, as on Mount Shasta and Mount Rainier, they become bigger and frequently spread to low levels. They gain the sea in 57° N. and at intervals for 700–800 miles (*c.* 1000–1160 km) northwards into Alaska. Here they attain their greatest breadth of 80–100 miles (*c.* 130–160 km) and an area roughly equal to that at America's other extremity. West of the big centres near Mounts Fairweather, Logan and St. Elias into the Alaskan Peninsula and the Aleutian Islands, the glaciers become smaller and narrower and less compact, since the mountains are lower and the north winds which blow over them in winter are cold and relatively dry. The snowline rises from 2000 or 2500 ft (*c.* 600–725 m) to 8000 or 10,000 ft (*c.* 2450–3050 m) in the Aleutian Islands, whose largest snowfields lie on Mount Makushian. The glaciers also decrease eastwards into the interior of Alaska which has valley glaciers only. These are small and their snouts rarely descend below 900–1200 m. Peaks of 1200 m, even north of the Arctic Circle, are frequently without snow.

Europe's irregular relief makes a similar comparison impossible. Yet Norway's glacierisation, which is the severest on that continent, becomes steadily less intense southward. Finally, small cirque glaciers only are to be found.

The following table which in the main is that compiled by Hess[278] gives the glacierised areas in the various regions. It departs from his estimates in substituting Åhlmann's figure for the Alps[279] (Hess, 3800 sq. km), K. J. Podesersky's figure for the Caucasus[280] (Hess, 1840 sq. km), his own later figure for central Asia,[281] S. Thorarinsson's figure for Iceland[282] for that of Thoroddsen (13,500 sq. km), N. N. Urvantzev's figure[283] for Severnaya Zemlya, F. Loewe's figure[284] for Greenland, J. J. Dozy's figure[285] for New Guinea, and H. P. Kosack's figure of 13,101,154 for the Antarctic ice (see p. 78). The larger numbers are all approximate.

Continent	*Region*	*Area (sq. km)*	
Europe	Pyrenees	40	9,180
	Alps	4,140	
	Norway	4,600	
	Sweden	400	
Asia	Caucasus	1,968	c. 100,000
	Other areas (Karakoram: 13,660; Himalaya 10,000)	100,000	
North America	Alaska	80,000	90,000
	Other areas	10,000	

Continent	*Region*	*Area (sq. km)*	
South America	—	25,000	25,000
Africa	—	20?	20?
Australasia	New Zealand	c. 1,000	1,015
	New Guinea	15	
Arctic	Novaya Zemlya	15,000	1,866,955
	Franz Josef Land	17,000	
	Spitsbergen (North-East Land, 11,225)	58,000	
	Iceland	11,785	
	Severnaya Zemlya	15,100	
	Jan Mayen	70	
	Greenland	1,650,000	
	North American islands	100,000	
Antarctic	Continent	13,101,154	13,104,000
	Islands	3,000	
		Approx. Total	15,193,000

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249 *493*, 66; *1840*, 187. **250** *767*, 45. **251** *1439*, 177. **252** R. Bentham, *G. J.* 97, 1941, 144. **253** *927*, 345. **254** *1729*, 21. **255** *1840*, 155. **256** *518*, 18. **257** *1742*, 469. **258** *1840*, 161. **259** *Ibid.* **260** *1455*, 770; E. Herrmann, *Z. Gl.* 19, 1931, 267. **261** *522*, 150; *1537*, 12; W. Behrmann, *Z. Gl.* 15, 1927, 136. **262** *1677*, 38. **263** *493*, 64, 73. **264** *Am. G.* 19, 1897, 265. **265** W. L. S. Fleming, *G. J.* 96, 1940, 93. **266** *730*, ch. 8; *748*, ch. 4. **267** *227*; *227* (2); K. Oestreich, *Z. G. E.* 1914, 435. **268** *152*, 52 (lit.). **269** *1750*. **270** *1341*, 10; *1343*, 131; G. Merzbacher, *Bayer. Ak. Wiss.* 1928 (8 sheets). **271** *494*, 25, 111; B. W. Tronow, cf. *G. Zbl.* 15, 1927, 77. **272** *753*, 362; J. Perthes, *Taschen Atlas der ganzen Welt*, Ed. 67, Gotha, 1931, 10, 16. **273** *1840*, 11, 134; cf. *J. Gl.* 2, 1954, 378. **274** *1092* (8), II, 892. **275** L. C. W. Bonacina, *A. int. Hydr. sc. B.* 23, 1938, 83. **276** *1435*, 303; *1444*, 32; C. K. Wentworth & D. M. Delo, *op. cit.* **277** O. D. v. Engeln, *G. S. A. B.* 41, 1930, 99. **278** *748*, 114; *753*, 362; *755*, 121. **279** *G. Ann.* 6, 1924, 251. **280** Cf. *Z. Gl.* 7, 1913, 287. **281** *755*, 103. **282** *1704*, 172; *G. Ann.* 25, 1943, 17. **283** Cf. H. W:son Ahlmann, *Y.* 53, 1933, 379. **284** *Btr. Gphys.* 46, 1936, 317. **285** *Z. Gl.* 26, 1937, 45.

Chapter V

ICE-MOTION

1. *Facts of Ice-motion*

Measurement. Glaciers, following the common opinion of the ancient Greeks,[1] were considered to be composed of rock crystal as late possibly as the beginning of the 18th century—mention may be made of the younger Pliny, J. Simler and J. v. Mural—yet few held this view after the middle of the 17th century. Though W. G. Ploucquet[2] in 1789 denied their onward movement, the fact that ice moved was early known to the Swiss peasants and was familiar to them and to a number of naturalists[3] towards the end of the 18th century because marks indicating favourable crossing places on glaciers were displaced and ice invaded ground previously uncovered. Ice-flow was probably known even earlier in Iceland for the word *skridjökull* ("moving glacier") is several hundred years old[4]: that ice moved was expressed in 1695 by T. Vidalin and in 1794 by S. Pálsson. Gruner,[5] in 1760, was apparently the first to give a measure of the actual velocity—a stone travelled 50 paces on the Grindelwald Glacier in six years. Methodical studies began with Hugi[6] who ascertained the rate from the shifting of his hut on the Lower Aar Glacier—the hut moved 330 m between 1827 and 1830 and 1428 m between 1827 and 1840. Between 1840 and 1922 the distance travelled was 2·858 miles (4·6 km) or an average of 184 ft (56 m) per annum.[7]

The simpler laws were first deduced in Switzerland where Europe's glaciers are most accessible, the subject of glacier-physics was first studied, and the nomenclature was developed: Forbes[8] on the Mer de Glace de Chamonix, Agassiz[9] on the Lower Aar Glacier, and the brothers Schlagintweit[10] on the Pasterze were the pioneers. The last decade of the 19th century saw research extended into the Arctic, chiefly Greenland—T. C. Chamberlin (1894), G. F. Wright (1894), R. D. Salisbury (1895), R. S. Tarr (1897)—and Alaska—H. F. Reid (1891, 1896) and I. C. Russell (1890, 1891). Nevertheless, the Alps, in which the focus of study spread eastwards, continued to provide the basis for most conclusions as formulated by A. Heim, E. Richter, E. Bluemke, S. Finsterwalder, H. Hess and later by H. Bader.

The earliest determinations, like some later ones, had more or less the nature of accidental discoveries: clothes lost on the Glacier des Bossons were subsequently extruded; the ladder which de Saussure's guides left in 1788 on the Mer de Glace was picked up in 1832 4050 m lower down,[11] Hugi's hut was shifted (see above); Agassiz's "Hôtel des Neuchâtelois" (or its remains) moved 4·6 km between 1840 and 1922[12] (see above); and bore-rods placed by Hess in the Hintereisferner in 1901 had moved 1090 m by 1931.[13] The first deliberate attempts were based upon the displacement of stones; H. Besson,[14] the first to use this method, placed them in a straight line across a glacier between two fixed points on its banks. Later, F. A. Forel[15] suggested that they should form a continuous row and Heim that they should be painted and placed in a series of lines. Yet stones give a false idea of velocity since they have their own motions.[16] "Tabling" and gliding are chiefly responsible

for this,[17] as is seen in Norway and on the Hintereisferner where the annual error totalled 2 cm, though thin stones melt their hollows downstream by pressing against their side.[18] Big boulders on steep slopes facing south have the highest proper motion. This is relatively large on glaciers whose annual flow is very small and reaches its maximum in summer towards the snout where melting is excessive. To overcome this, stones were anchored with feet of steel wire[19] or chosen of such a shape and size that they neither sank nor tabled[20] (see p. 61).

Another early method, used by A. Escher in 1841 on the Aletsch, was to drive stakes into the ice. More recently[21] the rods have been inserted into bores. Rods have been used on the Rhône Glacier since 1882 and are preferable to stones which, especially in the névé, are quickly buried. On the ice-sheets which have no stones or similar points of reference, positions are fixed by bamboos or ice-pinnacles,[22] as was early done in Greenland by A. Helland, R. Hammer, C. Ryder and K. Steenstrup. Dirt cones have also been used.

In measuring glaciers, three methods may be employed: the graphical, tachymetric or trigonometric.[23] A fourth or stereophotogrammetric method is increasingly being used[24]: photographs are taken from the air across the ice-current or with a photo-theodolite on the ground, the photographs being repeated from the same point at intervals of some days or by a pair of short-base photogrammetric cameras. The forward movement of stones or crevasses on the glacier-surface can then be measured with great accuracy by a stereo-comparator as parallaxes. The method has the advantages of accuracy and shortness of field work to commend it—it may therefore be used for mapping glaciers by short expeditions—as well as the ability to measure glaciers or parts of glaciers that are badly broken or crevassed, though it is less suitable to the uniform surfaces of névés. The measurements, being taken from pictures that are preserved, can be checked later at any time.

Exact measurements[25] have in these ways been made, for example, for the Alpine Obersulzbachferner, Vernagtferner, Hochjochferner, Hintereisferner, Gepatschferner, Guldenferner and Yamtalferner. In other parts of the world, relatively few glaciers have been similarly surveyed; Drygalski's work in west Greenland[26] and R. Finsterwalder's mapping[27] in central Asia (Pamir, Himalayas), the Cascade Range and South American Cordilleras are notable examples—other work has been done in Spitsbergen and on Mount Kenya.

The annual velocity may also be roughly gauged by the distance between transverse crevasses,[28] moulins[29] or dirt-bands (see p. 50). Lines on a map joining points of equal rates of movement, such as have been drawn for the tongue of the Pasterze[30] (fig. 29), have been named "isotachytes".[31]

Laws. Although we know the distribution of surface-velocity for comparatively few glaciers and the velocity in a vertical sense for still fewer, we know enough to arrive at qualitative laws. They resemble the laws governing the flow in rivers; the analogy between rivers and glaciers has, indeed, been often stressed,[32] the difference being mainly quantitative and due to the great viscosity of ice.

Longitudinal velocity, evinced by crevasses and dirt-bands, varies with the slope of the bed and of the ice-surface, with the depth, width and

cross-section of the ice, with the amount of associated water and with the temperature.[33] Mean velocity, surface-slope and cross-section are connected by Etyelwein's formula for the movement of water in rivers and canals.[34] Since the product of the velocity and cross-section is constant,[35] the flow in constrictions is accelerated, as on Engebrae, Norway.[36] The effect of thickness was studied mathematically by A. A. Odin[37] though Agassiz,[38] fo'lowing É. de Beaumont,[39] had already noted that hanging glaciers, though steep, flowed relatively slowly because their basal friction was great.

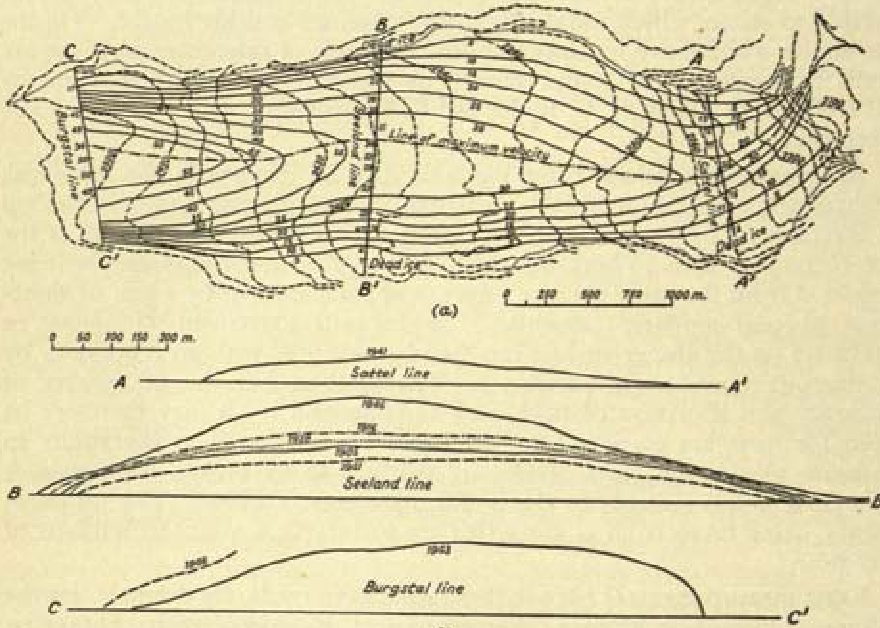


FIG. 29.—Isotachytes on the Pasterze tongue (*a*) and velocity diagrams along three transverse sections (*b*). V. Paschinger, *1253*, p. 51.

The velocity rises from the névé which, though little studied,[40] flows slowly owing to its breadth and shallowness and gentle slope to a certain point on the tongue. It diminishes thence with lessening thickness towards the snout[41] unless, as in the Mer de Glace and some Greenland glaciers, the gradient steepens. It rises towards the snout in those glaciers which pass into ice-fjords since the sea-water fills their cracks and increases their mobility and speed.[42]

Recent investigations,[43] following upon earlier observations,[44] suggest that the line of most rapid motion dives steeply downwards from the bergschrund into the firn basin (as Agassiz[45] was the first to notice), rises about the firn-line, and appears at the surface above the snout. This disposition of the flow explains the tilting of the stratification in the firn (fig. 30), the downward widening of crevasses in the firn (see p. 45), the verticality of bore-holes several years old when exposed to view again by the ablation of the newer layers during an excessively warm summer, the freedom from crevasses in the

firn of detritus, even near the bergschrund, and the grinding out of the rock-basins (see p. 274).

Observations[46] on the Pasterze and other glaciers confirm the law previously found on the Hintereisferner[47] that the summer movements, because of the lower frictional resistance caused by soaking the tongue and sole with water, are greater in the lower third while in the remainder, including the firn itself,[48] the winter velocity is higher because of firn pressure—the winter flow in the firn is of two types, one the movement of the glacier proper, the other which is superimposed upon it the creep and compaction of the snow.[49]

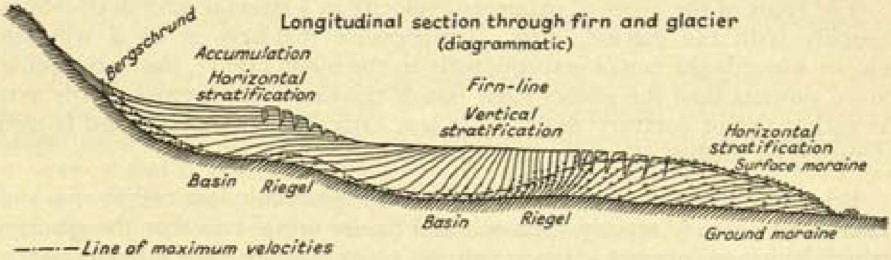


FIG. 30.—Course of stratification planes and lines of flow in a cirque-glacier. R. Streiff-Becker, *1649*, p. 15. fig. 9.

The statement, based upon careful measurements,[50] that flow takes place towards the source at night and towards the extremity during the day, and in two contiguous parts may be in opposite directions at the same time, does not apparently apply to all glaciers and may itself rest upon errors introduced by the temperature of the instruments.[51]

The components in composite glaciers may have very different velocities. Some may even be inert,[52] as in the Asau Glacier of the Caucasus where the middle flows between two long stretches of dead ice severed from their feeding grounds.

Contrary to the view of early writers[53] who were led astray by the direction of the marginal crevasses, the velocity rises inwards from the edge. A. Guyot[54] and R. Mallet,[55] the first to recognise this, appreciated the effect of marginal friction and the true significance of the crevasses and terminal moraines. Direct proof was given later by Rendu[56] and by measurements and stereophotogrammetry of certain Alpine glaciers[57] though moraines and marginal crevasses make them difficult. Exceptions occur where, as in the Fröya Glacier of north-east Greenland, a presumed buried ridge slows down the flow along the median line.[58]

That the curve is not a parabola, as Viollet-le-Duc thought, has often been demonstrated,[59] though it approximates to this curve on larger glaciers.[60] The rate rises very rapidly from the margin but remains fairly constant over the middle stretch, possibly as a result of the enhanced plasticity and therefore smaller erosion[61] (cf. p. 231). The increase, due to lessened friction at the bottom and sides and to great thickness,[62] varies with the shape and composition of the bank and with the cross-section. The curve is more flattened

in U- than in V-shaped valleys[63] and its critical points correspond to those in the cross-section of the bed as Forbes[64] anticipated and bores through the Hintereisferner have proved.[65] Retardation also rises with the gradient and seasonal temperature[66] and varies inversely with the total movement.[67] The 20 km broad Storglacier of Greenland which in its centre flows at 1700 m/annum has an annual marginal velocity of only 10–50 m.[68] The increase towards the centre where the flow is constant may be either linear or with a quadratic or still higher function depending upon the cross-section of the bed. In glaciers with a block movement (see p. 118) the rate of flow rises much more rapidly from bottom and sides,[69] and also of course not regularly.

The locus of the point of maximum velocity in a straight stretch coincides roughly with the glacier's axis. Ice-pressure displaces it in a winding glacier towards the bends (exceptionally to the inner side[70]), the curve being more sinuous than the glacier itself but in the same sense, crossing the axis at each point of contrary flexure. It was early demonstrated[71] and is well depicted by median moraines and dirt-bands.[72]

Below a confluence, the inner margins slope gradually less and the lines of maximum velocity rapidly approach and finally unite[73] so that the glaciers which had been separate entities coalesce, as is shown by flow measurements, by ogives and by the pattern of crevasses. The movement to the sides, which is partly responsible for spreading the moraines, has been proved on many glaciers[74] and most clearly by a long series of observations on the Rhône Glacier.[75]

Early observers,[76] despite the unreliability of their figures, were able to show that the flow in valley glaciers decreased downwards. The verticality of moulins, proved by Agassiz,[77] led Forbes[78] to affirm that the decrease was mainly nearly the bottom. This conclusion has been sustained[79] by measurements, by shafts, by the absence of strain on bore rods, and by calculations on the Hintereisferner. The bottom velocity on this glacier and on the Vernagtferner and Pasterze is 75% or even 95% of the surface velocity[80]; the fraction increases during a retreat and decreases during an advance and with quickened flow. The maximum velocity, it is said, may be below the surface as in rivers[81]; for bores in the Hintereisferner were constricted 8–12 m down.[82]

Figures of velocity gradient[83] demonstrate that the basal layers flow extremely slowly if there is no sliding, though striated surfaces prove some flow. The law governing the downward reduction of velocity has yet to be discovered (see pp. 114, 123).

Friction and clogging with debris reduce the basal flow,[84] especially at low temperatures. Debris, unequally distributed, causes local variations or even stagnation. Clean ice advancing upon consolidated gravels charges its basal layers and so flows more slowly. Hence, the distance to which a glacier descends is affected by the subjacent rocks: it is more for example on sandstones than on shales.[85]

This downward decrease of flow, accepted as almost axiomatic from the early observations on mountain glaciers and from the analogy with rivers, is not universally true, viz. in those parts of valley glaciers where the flow is

over level ground or through rock-basins or in the case of ice-sheets or plateau glaciers (see p. 123).

The "brittle" upper crust, with its crevasses, rides more or less passively on the plastic ice though with some motion of its own. Flow is not by "gravity or drainage flow", as it is in those cases where the velocity decreases downwards, but by an "extrusion or pressure flow" (see p. 122).

Early investigators neglected vertical movements and those they did determine were unreliable. The first extended series of these difficult measurements was made on the Karajak Glacier, west Greenland.[86] Upper movements towards the margin were observed by Agassiz[87] and later by glacialists[88] on the Hintereisferner and the glaciers of Greenland and of Yakutat Bay. They are confirmed by the raised edges[89] at the contact of two glaciers, on the concavities of bends, and where morainic protection is lacking.

The obliquely outcropping ice-layers which arise from differential melting between the outcrops of dirt-bands form a series of small, parallel terraces or ridges concentric with the snouts.[90] The upturning varies from a few degrees to vertical and is commonly 30–40°. It is more at the end than at the sides and is conspicuous to considerable distances from the snout.[91] This swelling and upturning towards the snout, with its accompaniment of thrust planes, arises in several ways,[92] e.g. from the action of the glacier-bed and of basal debris, from obstructions of schotters, moraines or snow-bank glaciers, or from melt-waters freezing in these places as the result of the decrease of pressure. It is satisfactorily explained by the movement of stationary glaciers.[93]

Observations show that the flow varies seasonally[94] (contrary to an occasional statement[95]) because of the penetration of summer heat and seasonal melt-waters. The ratio of summer to winter flow, which has been given as 2:1 (Tyndall) or 9:8 (J. Vallot) and is often 4:1, is highest towards the snout and in thin and small glaciers or in those which attain the lowest levels. In large valley glaciers the variation is below 10% of the velocity.[96] Hence Greenland glaciers, notably if they issue from the ice-sheet, vary little if at all since their great depth and low temperatures are inimical.[97] A review of the literature[98] shows that a big summer advance characterises advancing or stationary glaciers and is related to the small difference between the surface and bottom velocities.

Since the temperature and amount of water are inconstant and vary not a little with the season and from year to year, glaciers vary in their flow over a number of years. This was conclusively shown for the Vernagtferner[99] though the variations, as elsewhere, are less near the margins. The same factors are responsible for the quicker flow during the day, as proved for instance on the Pasterze[100]: Agassiz[101] erroneously entertained the contrary view.

Data. Although the velocity at a particular point is subject to strong variations and, as on the Crillon Glacier, Alaska,[102] changes from day to day so that measurements confined to a few days are unreliable guides,[103] we know enough to gain some idea of the velocity in various parts of a glacier and at various periods. Its annual amount in the bigger Alpine glaciers[104] is between 30 m and 150 m, or perhaps one foot (30·5 cm) a day: in the ice-falls of the eastern Alps it is 20–28 cm.[105] Norwegian figures are similar[106] and may reach about 35 m. This is true too of Spitsbergen glaciers[107] which are

relatively thin and have much basal debris. Few satisfactory data are available for Iceland, though O. Torell's observations appeared as early as 1857[108]—Hoffellsjökull moves at an annual rate of 225 m, 1400 m from the ice-front, and may have a maximum annual speed of 700 m.[109] In Greenland, on the other hand, much work has been done in this connexion since Helland's first determination in 1875[110] and the somewhat later ones by Steenstrup and Kornerup.[111]

Helland's figure of 20 m/day on the Jakobshavn Glacier, though deemed untrustworthy,[112] has been fully borne out by even higher rates[113] and particularly by Drygalski's extended series of measurements on the Karajak Glacier[114] (70° 23′ N. Lat.). Nevertheless, these observations, which give daily values of 15–30 m, are exceptional[115] since the border of the ice-sheet has an average flow of less than 30 cm/diem[116] and glaciers from the island-ice flow at a rate comparable with the Alpine.[117] High velocities distinguish only the "running" or outlet glaciers which are pressed into fjords, notably ice-fjords transecting the mountain barrier, and because of the diminished resistance which floating, calving and bathing of the ice in water occasion, move more rapidly than glaciers which end on land.[118]

The flow lessens from the margin into the interior of the ice-sheet[119]; the daily rate of 18 m on the Upernavik Glacier was only 2·5 cm on the ice-sheet a little distance back from the margin. The ice-sheet's annual velocity[120] has been given as about 11 m or 22 m.

Data suggest that the daily average rate of the Himalayan and Karakoram glaciers is not less than 50 m[121] though in the Rakhiot Glacier of Nanga Parbat it amounted to 130–800 m/annum.[122]

The Antarctic ice in regard to the velocities within its mass and throughout the year probably varies less than any other ice.[123] It flows more slowly than Greenland ice as L. Bernacchi, J. G. Andersson, R. F. Scott, W. S. Bruce and E. v. Drygalski have shown—paucity of melt-water, extreme cold and small plasticity, combined with relative thinness and the attainment of the coast over almost all its length, account for the difference[124]; the shelf-ice and the Antarctic bergs calved from it have a high degree of rigidity and show no signs of plastic deformation.[125] Its slowness is strikingly exemplified in the inability of the Ferrar, Mackay and Fry glaciers to raise pressure ridges in the sea-ice about them,[126] as well as in the following figures[127]: the Beardmore Glacier, in its swiftest part, flows *c.* 90 cm (3 ft)/diem; the Ferrar Glacier, 4·5 cm; the Mackay Ice Tongue 84 cm; and the ice at Gaussberg 40 cm, becoming less inwards from the margin. The annual velocity of the Barne Glacier is *c.* 9 m and of the Blue Glacier, Royal Society Range, *c.* 1·2 m. Considerations of precipitation, ablation and flow suggest that the snow in the interior of the Antarctic may reach the margin only after 20,000, 25,000 or even 50,000 years[128] (cf. Greenland, p. 151). Corresponding figures for the Rakhiot Glacier 15 km long and 200 m average velocity are 75 years, for the Fedchenko Glacier, 1000 years,[129] At least 500 years have been assigned as the age of the snout of the Lower Aar Glacier.[130]

Knowledge of the depth and velocity of a glacier enable calculations to be made of the volume of ice that passes annually across a particular profile. The Fedchenko Glacier, for example, which has an annual velocity of 170 m and an inclination of 4% and a thickness of 550 m in the middle part of its long stream, was crossed here by *c.* 111 million cu. m of ice.[131]

2. *Ice-temperatures*

Having obtained some idea of the flow of glaciers we have now to consider the vital question of its cause and the preliminary question of the distribution of temperature within the ice.

Figures of internal temperatures are scanty and often of a limited usefulness; for moulins and crevasses are affected by air-currents and like bores are generally shallow. Hugi on the Grindelwald and Agassiz on the Lower Aar Glacier made the first investigations.[132] Indeed, until the revival of interest in the subject towards the end of the 19th century, their results stood almost alone.

Glaciers consist of two zones, an outer one of fluctuating temperature and an inner one in which the temperature is approximately uniform. In the former, the ice behaves like any other rock and its temperature is susceptible to oscillations governed by the amount of snow-cover and by the changing temperature of the air above from hour to hour, day to day, season to season, and year to year.[133] The change is effected by conduction and by air-circulation and is felt in Greenland to a depth of 50 m.[134] Agassiz found that the daily oscillation died out 2–5 m down in bores sealed at the surface and that between 30 m and 60 m there reigned a constant temperature of 0°C, a result others had already anticipated from theoretical studies.[135]

Each day, notably a foehn day, sends a wave of heat-energy down into the glacier, each night a wave of reverse nature. These weak waves penetrate quickly and with deceasing amplitude. They vanish completely at a shallow depth[136] which on the Jungfraujoch is 10 cm and in Greenland's interior is 1 m and nearer its margin 0·7–0·8 m. Each season sends greater and deeper waves whose penetration depends upon the difference between winter and summer air temperatures and the consistency of the ice: the more uniform the surface temperatures, the deeper is the layer of stationary temperature. The amplitude decreases exponentially downwards[137] (fig. 31): in Greenland it amounted at the surface to 40°C and at a depth of 20 m to only 0°1C.[138] The cold maximum of the long winter wave enters later as it marches downwards and its velocity which in Greenland was 1 m/month[139] is reduced in the same sense. The waves, for example, required two days to pierce 1 m and ten days to penetrate 2 m.[140] Cold waves travel more slowly than warm waves in which percolating waters aid conduction[141]: the disturbance of the firn by digging, which by destroying the ice-bands and layers of low permeability (see p. 32) allows the melt-waters to penetrate more readily, explains why shallow pits give higher temperatures than bores.[142]

The temperature of the variation layer, due to interference of cold and warm waves, oscillates between 0°C and the normal air temperature of the locality.[143] The temperature-depth curve on Isachsen's Plateau in West Spitsbergen, and in the Jungfraujoch, is in fair agreement with that calculated from the known air temperatures and the thermal constants of the firn.[144] The surface temperature at night and in winter in Alpine glaciers is depressed below 0°C but affects only one or two metres. Where the cold is more or less permanent, the upper strata are always below 0°C; the névé of Mont Blanc is constantly at −16·8°C at a depth of 10–13 m[145]; the Antarctic gave a reading of −29·4°C[146]; and seismic methods 120 km from the ice-edge in Greenland yielded a temperature of −16°C at a depth of 180 m.[147]

The depth reached by seasonal waves is much less than the 50 m Heim[148]

assigned to it. Representative figures[149] are 1 or 2 m in west Greenland, 14 m on the Borgfjord, Greenland, and 15 m in central Greenland; 12–15 m on Upper Seward Glacier, Yukon Territory; 15 m on the Tête Rousse and Jungfrau; 14 m on Mont Blanc; 8 m on the Hintereisferner; and 1 m on Snow Hill, Antarctica.

The zone of constant temperature is present everywhere, unless the ice, as around the edges, is thinner than the zone of fluctuating temperature. Its

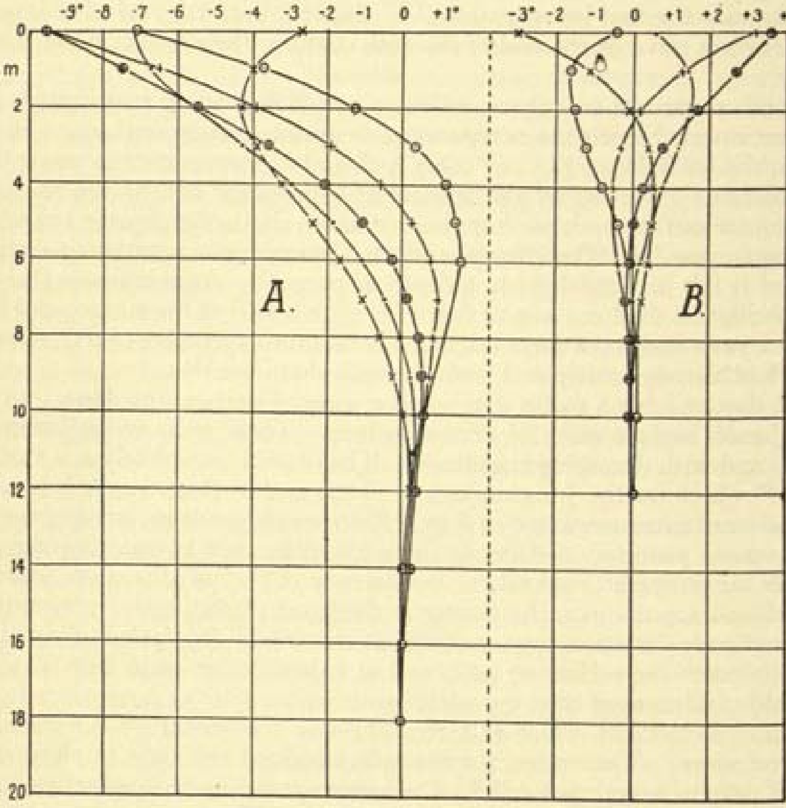


FIG. 31.—Yearly and seasonal waves of temperature in Greenland ice. Whole year (A), half-year (B). ⊙ December, + January, ⊕ February, · March, × April. 926, p. 224, fig. 114.

highest layer has the mean annual temperature of the air. Where, as generally in Scandinavia and the Alps, this is not far from 0°C, the glacier is approximately at freezing point.[150] If, as in high latitudes, the annual temperature is far below 0°C, the glaciers are "cold".[151] Thus this horizon in Greenland ranges between −15°C and −32°C.[152] The Antarctic has given comparable results[153]; the large temperature variations at Maudheim, which ranged from 0°C in summer to *c.* −45°C in winter, were quickly damped down so that at a depth of 5 m the swing was only about 4°C, at 10 m 1°C, and at 20 m only 0·05°C—the mean annual temperature at a depth of 100 m was −16·2°C and at 185 m was that of freezing point of sea-water. Glaciers in the Mount Everest region are frozen throughout.[154] The prevailing characteristics of these "cold" glaciers are relevant to their

frozen state[155]: basal melt-water is absent; repeated imbricate faults, shear planes and drag folds traverse the basal layers; terminal and side walls are vertical or overhanging; composite glaciers have discordant and overriding junctions; and many glaciers adhere to steep mountain sides.

J. P. Koch[156] found that the temperature at greater depths rose 1°C for every 20 m. Drygalski's serial observations,[157] the first to give trustworthy information about the temperature throughout the year in a glacier's mass, showed that it approached the melting point and that the Greenland ice had this temperature below a certain horizon. Koch's conclusion[158] that this is at least 140 m down agrees with observations[159] in the Antarctic and in North-East Land where the temperature was negative to at least 150 m and probably to the base at 270 m. The isotherm of 0°C lies beneath the Barnes Ice Cap of Baffin Land[160]—no water issues from beneath the ice. Bores in

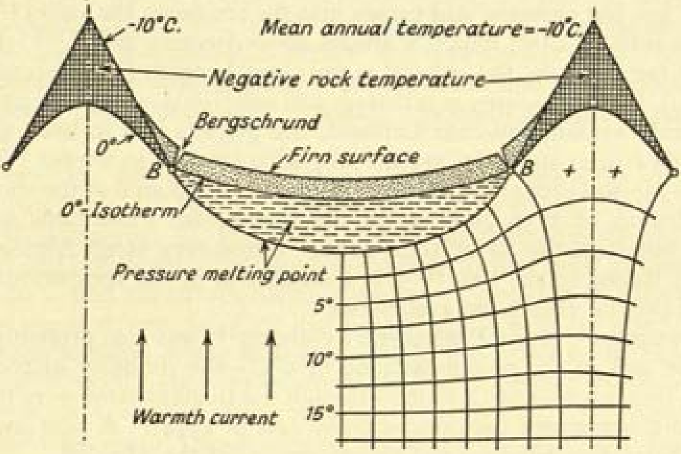


FIG. 32.—Temperature conditions in a firn basin and its rock-bed. R. Haefeli & P. Kassler, *A. int. Hydr. Sc.* 1948, II, p. 312, fig. 6.

the Hintereisferner[161] showed that the temperature is that of the freezing point corresponding to the pressure-conditions at each depth. This was confirmed on the Jungfraujoch[162] and follows too from a consideration of the factors known to be at work within a glacier[163] according to the Clausius-Clapeyron equation and including the heat generated by friction,[164] as well as from experiments of the effect of pressure on ice-temperatures.[165] Hess[166] affirms that the difference between the observed and calculated temperatures increases downwards but that the results are not accurate enough to justify the conclusion that it represents the effect of lateral pressure due to flow. Thick glaciers, it has been suggested,[167] do not attain the critical pressure-temperature, the actual temperature being below it.

The basal layer, since the days of Hugi and G. Bischof, has been thought to be roughly at melting point[168] and slightly below 0°C. Exceptions occur where the ice is thin and free from snow, as around nunataks in east and south Greenland[169] and along the margins of glaciers and névés[170] (fig. 32) and in the glaciers of the high polar type of north-east Greenland where the ice even in summer is frozen throughout its mass.[171] There are several reasons for this view: bores contain water in summer even down to the sole of the

Hintereisferner[172] and to a depth of *c.* 21 m in North-East Land[173]; subglacial streams flow in winter[174] (see p. 67); a considerable volume of subglacial water is essential to prevent the temperature falling below melting point[175]; and there is a continuous and uniform supply of heat from below (this is 100 times that due to flowage[176]) and from friction along shear planes.[177] Considerations of the geothermal gradient lead to the conclusion that the gradient in ice is less than in rock and that a part of the heat is used for basal melting.[178] Seismic methods suggest that even at the base of the Greenland ice-sheet large lakes of water are dammed up.[179] Hence, contrary to the early view,[180] a glacier, except near its bergschrund, is not frozen on to its rocky bed. Nevertheless, glacier-lakes, striated surfaces and bores prove that the contact is close.[181]

The nature of the contact is intimately connected with the isogeotherms. In an average glacier, the melting point isotherm is probably in the ground below the firn but emerges and passes into the ice down the valley[182] so that the tongue rests in a bed which is always above freezing point.[183] In Spitsbergen,[184] for example, the ground is unfrozen beneath the ice (except near the margin), bottom melting is initiated, and subglacial streams begin to flow. In the Sareks region, Swedish Lapland, the glaciers rest in beds which are permanently frozen and there are no subglacial streams in winter.[185]

The annual isotherm of 0°C of the Alps lies in the ground at the snowline[186] or at elevations of 2750–2000 m (in the Karakoram Mountains somewhat lower[187]) but is in the air at 1900–2000. Some very steep Alpine glaciers are clearly frozen to the base[188] and a small rise in the temperature of the subglacial ground causes ice-avalanches.

The temperature of the Pleistocene ice-sheets[189] was also probably highest at the base and in the neighbourhood of 0°C—the mobility of the ice was facilitated by saturation with water—though the isogeotherms were depressed and the cold penetrated the ground below (see p. 1344). A cold layer 100 m thick has been postulated for the Pleistocene ice of the Alps.[190]

These considerations lead naturally to the idea of a geographical classification of glaciers,[191] viz. into "temperate" (Åhlmann) or "warm" (Lagally) glaciers and polar glaciers, the latter subdivided into highpolar and subpolar or "transition" (Lagally) glaciers. In temperate glaciers, e.g. in Europe, Spitsbergen and North-East Land, the temperature is at melting point in winter and the top layer has a negative temperature to a depth of a couple of metres (see p. 105); polar glaciers have a negative temperature even in summer to a depth of 100 m; the high polar type has no thaw water; and the subpolar type has thaw water in the accumulation area in summer.

3. *Theories of Ice-motion*

General. Ever since de Saussure made his famous ascent of Mont Blanc in 1787 and began his glacier researches, glacier-motion has been a topic of lively interest and debate. J. J. Scheuchzer and J. H. Hottinger were the earliest to speculate upon the cause though L. Bordier made the first sound guess by comparing glacier-ice with wax. Playfair[192] imagined that glaciers were undermined by earth's heat and impelled by their own enormous weight. Altogether about 80 theories of glacier-movement have been propounded.[193] In discussing the most important of these a mathematical treatment has been deliberately avoided.

Ice-motion, according to the laws of fluid mechanics, is controlled by the gradient, the volume and depth of the ice, and the potential energy not expended in overcoming internal cohesion and basal and marginal friction. What, however, is the physical cause? Ice moves as a plastic substance, yet possesses crystalline structure. How are the two to be reconciled? Numerous geologists and physicists have conducted this research.

Although the subject is more and more being placed upon a rigorous scientific basis, a complete explanation is still awaited. Many of the hypotheses outlined below contain elements of truth and a complete theory should take all into account. None, however, is adequate alone and most involve assumptions that do not command general assent.

Experimental support of the various theories is incomplete and unconvincing. Systematic observations carried on for decades, such as those made since 1874 on the Rhône Glacier[194] and for shorter periods on some Tyrolese glaciers, e.g. the Hintereisferner,[195] are required since accurate deduction is only possible on a foundation of precise and quantitative facts. Knowledge of the velocity's distribution along various profiles and throughout a glacier as well as of its economy are essential. Physical constants are likewise important.

Structure and physical constants. P. Groth[196] thought ice possessed trigonal symmetry. X-ray examination, however, seems to confirm the older view of W. Scoresby, G. Hellmann and E. Belcher that its symmetry is hexagonal and holohedral,[197] though a cubic form may exist at temperatures below $-70°$C and under special conditions. The views therefore that ice was dihexagonal pyramidal[198] or had hemimorphic or polar axes due to the asymmetrical location of the hydrogen ions in the lattice with respect to the basal plane[199] would seem to be erroneous since neither piezo- nor pyro-electrical effects have been observed in ice[200] (such polar properties might be masked by twinning). Nevertheless, H. Bader[201] concludes from asymmetrical air bubbles in ice that the ice is hemimorphic. The space-group is D_6^4. There are four molecules of water in the unit cell of the structure, the dimensions of which are $a = 4{\cdot}53$ Å, $c = 7{\cdot}41$ Å. The ratio of the axes is $c{:}a = 1{\cdot}634$. The Laue-diagram of the ice and the arrangement of the atoms (according to G. Aminoff) are given in the accompanying figures[202] (see figs. 33 (a) and 33 (b)). The oxygen atoms lie in puckered hexagonal layers in which they are raised and lowered alternately. Adjacent layers are mirror images, and the sizes of the parameters are such that each atom is surrounded by four oxygen atoms in regular tetrahedral arrangement at a distance of 2·76 Å.

Dobrowolski[203] has given a complete review of the crystal.

The values of the two rays are 1·31041 and 1·30911,[204] figures differing only in the second place of decimal from those obtained by Bertin.[205] Sections normal to the principal axis are sometimes not isotropic[206]; this is due to strain or because the plates are not quite parallel or contain foreign elements.

The coefficients of linear expansion were determined by, among others, J. Dewar, E. L. Nichols, and A. J. H. Vincent: the values range between 0·000027 and 0·000054. The two coefficients recognisable in the grains are not detectable in the mass since the crystals are irregularly orientated. Conductivity is 0·0052–0·00568.

The physical constants at different temperatures are given in the Smithsonian Physical Tables,[207] by Hess,[208] and by H. Landolt and R. Bornstein.[209]

The cohesion of ice is 7–8 kg/cm; its crushing strength is 25 kg/sq. cm.[210]

Figures of the viscosity or coefficient of internal friction (μ) given by laboratory experiments vary between 10^{10} g cm^{-1} sec^{-1} and 10^{15} g cm^{-1} sec^{-1} according to the pressure, temperature and crystalline structure.[211] They agree roughly with those obtained from glaciers themselves, as from bores in the Hintereisferner which gave M. Lagally 10^{14} g cm^{-1} sec^{-1} and H. Hess[212]

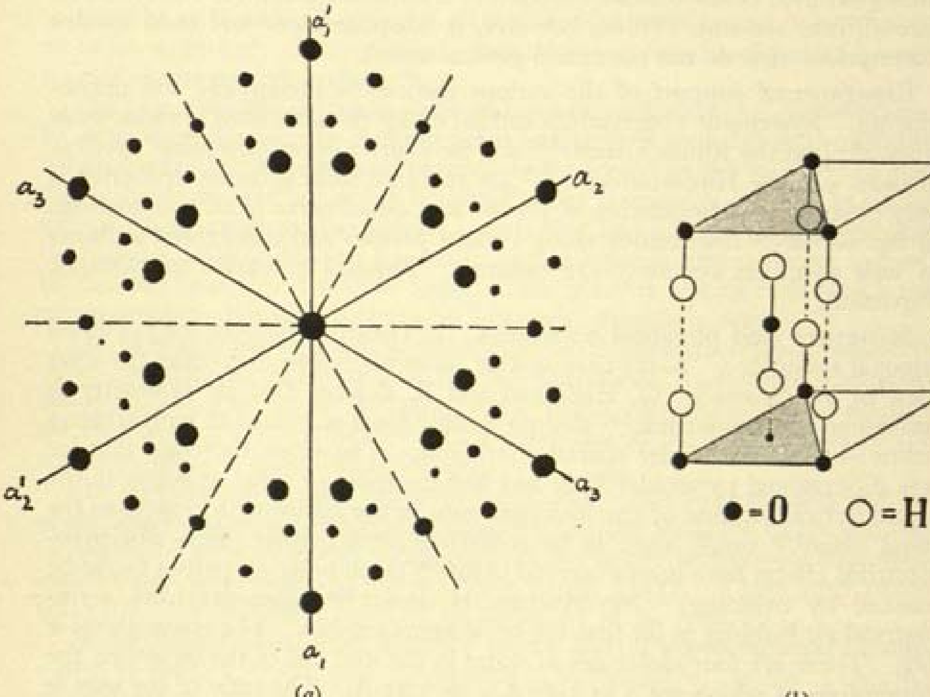


FIG. 33.—(*a*) Laue-diagram of ice (After H. Hess). (*b*) The movement of the atoms in an ice-crystal (After G. Aminoff). *755*, p. 2, figs. 1 and 2.

a figure at the base of $1{\cdot}8$–$2{\cdot}26 \times 10^{15}$ g cm^{-1} sec^{-1}. These figures may be compared with the following calculations[213]: Finsteraar $7{\cdot}2\times10^{11}$, Aletsch $3{\cdot}6\times10^{11}$, Mer de Glace $1{\cdot}9\times10^{11}$, Rhône $3{\cdot}5\times10^{11}$, Hintereisferner $5{\cdot}0\times10^{11}$, Fedchenko $4{\cdot}0\times10^{11}$, Notgemeinschaft $3{\cdot}8\times10^{11}$ and Zemu $3{\cdot}9\times10^{11}$.

The modulus of elasticity,[214] which ranges according to laboratory experiments between 92,700 kg/sq. cm and 23,632 kg/sq. cm and varies with the relation of the optic axis to the direction of force, has been seismically proved on the Pasterze to be 71,000 kg/sq. cm.[215] Ice yields elastically when pressures are slowly and carefully applied and is only permanently deformed beyond a certain minimum stress.[216]

Sliding hypothesis. One of the oldest hypotheses[217] attributes movement to a whole glacier sliding over its bed like a rigid body devoid of flexibility and plasticity. While de Saussure[218] stressed the importance of lubrication by subglacial waters, De Luc[219] thought glaciers were supported on

ice-pillars which gave way and W. Hopkins[220] arrived experimentally at the conclusion that they slide by sections or strips of indefinite width, parted by crevasses. The hypothesis afterwards found few adherents[221] for the arguments against it are weighty[222]; it ignores, for instance, differential flow and the displacement of particles at a rate depending upon stress and viscosity.

Nevertheless, some sliding almost certainly occurs[223] though it is difficult to evaluate since little is known of the coefficient of friction of a glacier upon its bed. R. M. Deeley and P. H. Parr,[224] who attempted to ascertain this by finding the amount of slip which would bring the actual surface velocity into agreement with the theoretical velocity, estimated the sliding for the middle of the Hintereisferner at 52·2% of the total movement. Hess calculated it for the same glacier to be at least two-thirds of the total surface flow and M. F. Perutz that half the surface flow of the Jungfraufirn was due to sliding on the bed.[225] Observations on the Rhône Glacier gave a figure of $4{\cdot}7\pm0{\cdot}2$ m/annum at the very snout.[226] The amount doubtless depends upon the gradient, the quantity of debris and the nature of the bed[227]; if this is rough and uneven, friction is great and sliding small; if polished and regular and well lubricated by subglacial waters, gliding may be not inconsiderable. While striated surfaces prove gliding,[228] the clear subglacial streams of winter indicate little at this season.[229] It is probably greater in periods of advance and feeble or wanting during retreat.[230]

Huge slices of ice not infrequently break off in the Alps.[231] For example, a mass 300 m long slid from the upper Arcelin Glacier[232] and 1 km long from the Glacier du Tour.[233] They are also known from other glacier-centres.[234] Probably the biggest example of the kind was furnished by the steeply perched Falling Glacier of Alaska which with a length of *c.* 1 mile (1·6 km) entirely slid out of its valley in 1905.[235] Such masses grade into ice-avalanches (see p. 22).

F. M. Stapf[236] thought the Scandinavian ice in north Germany glided over its ground-moraine, lubricated by melt-waters, and J. P. Koch[237] assumed a similar gliding for ice-sheets in general.

According to a related hypothesis,[238] the ice moved on saturated mud or on liquid rollers or buoyant cushions provided by subglacial streams and pools of water[239] derived from earth's heat, from springs, or from pressure-liquefaction. Only one-half of the sole of the Pleistocene ice, it has been said, was at any one time in contact with rock.[240] Cross-striae, intercrossings of erratics and oscillations of ice-sheets have been attributed to variations in the volume of subglacial waters.[241]

Dilatation. T. de Charpentier's dilatation hypothesis[242] (it had been anticipated much earlier[243]) imagined that waters from condensation or surface melting passed into the crevasses where they froze and by dilating drove the ice forward. This view found little support[244] except in the modified infiltration hypothesis which substituted capillaries for crevasses and made granular growth the expansive force.[245] In either form it is to be rejected[246] because it is inadequate and the ice-motion decreases towards the snout. Moreover, while water percolates by crevasses and moulins and to a depth of *c.* 2 m by capillaries since the *trous méridiens* drain during the night (in west Greenland, waters percolate 5 cm in 12 hours[247]) penetration of sound ice, though frequently affirmed,[248] is extremely doubtful.[249]

Alternate thermal expansion and contraction in the direction of easiest

movement, i.e. down the slope, has also been made the cause of steady progress,[250] as has linear dilatation induced by greater mobility due to rise of temperature.[251] This "crawling theory" illustrates the dangers of experiment *en petit*—H. Moseley,[252] its chief exponent, experimented on sheets of lead. It is manifestly fallacious[253]; for the glacier's internal temperature is practically constant and its advance is neither proportional to its length nor absent in winter when snows protect it. Moreover, heat melts the ice: it does not dilate it.

According to Croll's "molecular theory",[254] the heat received from sun's rays, mild rains, wind and percolating waters, propels the ice by weakening its cohesion and causing molecular motion. Each molecule descends step by step as it melts and solidifies. In the liquid state it occupies less space and so leaves room to move under the force of gravity to a lower level where it solidifies. But flow occurs in the polar night and in thick masses as quickly as in thin and is affected by slope.

Plastic theory. Forbes[255] ascribed the motion to plastic, viscous yield. Other naturalists had already anticipated this conclusion[256] or had noticed a glacier's resemblance to wax or flowing pitch.[257] Godeffroy[258] advocated a rolling over motion as in a lava, a view akin to that more recently put forward.[259] The movement has also been compared to that of solidifying plutonic masses or viscous lava-streams.[260] The word for glaciers in Norway (*brae* = paste or pulp) is in this respect interesting: *Kees* in Kärnten and Pinzgau is, however, not from German *Käse* (cheese) but from *ches* and High German *Eis*.[261]

The terms viscosity and plasticity have been used in discussions to denote the same physical property, although a viscous body is one which undergoes continuous permanent change of form under stress however small and a plastic body changes its form when stress exceeds a certain value—until the limiting stress is reached the body is elastic. While unconsolidated névé is pseudoviscous, ice in the main behaves as a plastic substance: its law of deformation is not a linear function between the shear stress and the rate of shear strain.

Cohesion is a property of rigid bodies, internal friction a property of fluids. That ice has cohesion is seen in crevasses and in the Chinese walls (see p. 35). It is largest at low temperatures and is reduced near melting point, especially under pressure when ice becomes plastic. Thus the question of cohesion and internal friction is one of ice-temperatures and these largely control the possibility and degree of movement.

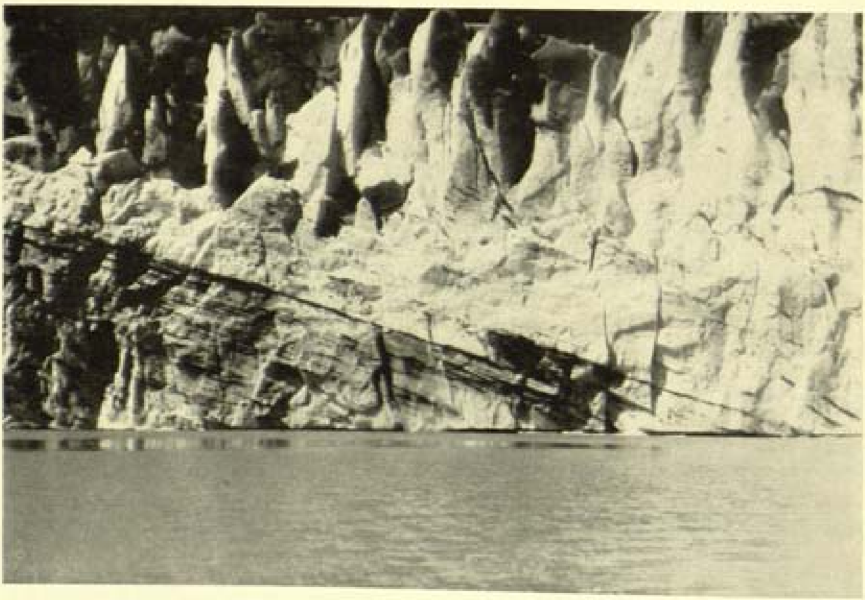
Glaciers behave as both rigid and plastic bodies. Their rigidity is seen in the scoring of rocks; in the pressure-waves produced where outlet glaciers join Ross Barrier; in the longitudinal folds sometimes induced where a glacier is constricted within a narrow, steep-sided valley[262]; in the shearing and buckling of the ice in its terminal part; in the backward slope of the ice-surface on rising ground, as on the impact side of nunataks or at the snout,[263] and in the observed occurrence of small isolated gaps (up to 3 cm in width) between the rock-bed and a cirque-glacier at a depth of 50 m.[264]

Plasticity is evinced by differential motion, by the disposition of dirt-bands, and by contortions and fluidal structures resembling those of schist or gneiss. Glaciers level up along their plane of contact, envelop boulders and obstacles, swell and sink around nunataks, narrow and widen along a valley, expand on

A. Ice-sheet, Dronning Maud Land, with nunataks
[Norwegian-British-Swedish Antarctic Expedition 1949-1952]

B. Ice-shelf, Ross Barrier near Maudheim
[Norwegian-British-Swedish Antarctic Expedition 1949-1952]

A. Air photograph of plastic folding on the Malaspina Glacier
[Harvard-Dartmouth Expedition 1933-1934]



B. Clean cut thrusts underlain by schuppen structure, Crillon ice-wall
[Harvard-Dartmouth Expedition 1933-1934]

(*facing p.* 113)

debouching on a plain, and in their external form adapt their mass to the topography. They form depressions in the lee of projecting spurs, insinuate themselves into fissures or the channels of subglacial streams,[265] flow occasionally over steps without crevassing,[266] groove their soles in the lee of boulders[267] and contract the cross-sections of tunnels dug into them.[268] Moreover, striae diverge round roches moutonnées, flute vertical or overhung rocks,[269] conform to twisted surfaces and tortuous channels,[270] and score concave faces of boulders, such as tabular flints in the Chalky Boulder-clay of England or fossils loose in the drift[271] (pl. VIA, facing p. 113).

Experiments on unit crystals and on granular aggregates of ice and the reproduction of almost all glacier characteristics on models of wax, mixtures of glue and plaster of Paris, or of pitch, the "poissiers" of W. J. Sollas, also bear out this conclusion. These laboratory experiments,[272] though unable to reproduce the glacier's mass and complicated environment, throw much light upon the problem of flow and its effects: they illustrate, for instance, the ascending currents in the lower layers when approaching an obstacle, the flow over loose material without its removal, differential movement, and the origin of dirt-bands and crevasses.

It is not surprising in face of all this and notwithstanding the elasticity and the granular structure and banding of the ice, that many glacialists[273] and probably the majority of glacialists to-day have favoured Forbes's view or contend that a glacier, if not truly viscous in the sense of Newton, Stokes or Navier and "creeps", is "viscoid", and in its "solid flow" is analogous to cold-rolled metals. The plasticity has been calculated[274] from data obtained on the Hintereisferner and determined by experiments.[275] These show that plasticity is not a constant but varies with the general structure, the coefficient in the individual grain being insignificantly less than that of the mass which is probably about 250 times that of pitch. The inner friction diminishes with pressure—Hess[276] dissented from this view.

That ice behaves like a highly viscous liquid and conforms to the laws of plastic substances is not to be denied: the aggregate flow of temperate glaciers at least simulates viscous flow. But we may doubt whether the flow has its origin in a true plasticity. Determinations of the coefficient of viscosity, based upon experiment and field observation, show considerable discrepancies, indicating variations according to the physical conditions. Thus plasticity is favoured by complete saturation and freedom from rock-debris.[277]

The cause of the apparent plasticity is somewhat conjectural. It is connected with temperature, stress, size, shape and orientation of the grains and is attributed to the following: plastic deformation of the individual grain[278]; translation along the lamellae[279] ("microplasticity" of A. Renaud); intergranular movement[280] ("mesoplasticity" of A. Renaud) facilitated by lubrication of "foam-cell" films[281] of saline solution or by intergranular and interlamellar water—ice behaves therefore as a two-phase system as far as stresses are concerned and obeys both thermodynamic and hydrodynamic laws[282]; growth of one grain at the expense of another[283]; and pressure melting and granular regelation.[284] Deeley's mechanical theory[285] ascribed it to viscous shear between adjacent grains and along laminae and to plastic shear of the granules into separate pieces by interchange of molecules from grain to grain. The seeming plasticity doubtless springs from these various factors and is conditioned by the grain and the intergranular and interlamellar

movements, facilitated by melt-water and regelation. *Fallwärme*, that is the rise of temperature due solely to the glacier's passage from higher and therefore colder altitudes to lower and warmer altitudes, may be a cause of melting and of flow.[286]

Fundamentally, the quasi-viscous flow of ice is due to local re-arrangements of atoms on the crystal lattice activated by their thermal energy and leading either to deformation of individual crystals or to the transfer of atoms across crystal boundaries. It is merely one practical instance on a gigantic scale of plastic deformation in a polycrystalline material subjected to sustained stresses.[287]

Experiment[288] has frequently demonstrated that plasticity rises with temperature and that ice at melting point is plastic without fracture: Andrew's curve for the relative hardness of ice at different temperatures shows this very clearly.[289] Since the upper layers, as in Greenland and Antarctica, are much below freezing point (see p. 105), they constitute a carapace of brittle, crevassed and more or less rigid ice. Their motion is induced by the lower uncrevassed layers[290] in the zone of flow where plasticity and therefore fluidity are augmented by the higher temperatures and the vertical pressure of the superincumbent mass. The lower layers[291] have the highest "relative velocity" but the least "absolute velocity", the surface velocity being the sum of the velocities of all the layers beneath. Yet E. v. Drygalski and others assert that the absolute velocity is highest in the lowest layers[292] because of the higher plasticity, pressure and temperature—the rate of flow may increase as the square of the depth.[293]

The actual base may move very slowly since experiments show that debris greatly raises the coefficient of internal friction.[294] The effect of temperature on plasticity is also seen in the more rapid flow of the northern glaciers of South Victoria Land[295] compared with those farther south, in the quicker flow of the arctic glaciers in summer and the relatively sluggish flow in Himalayan glaciers which have low temperatures and greater rigidity.[296]

The depth at which in temperate glaciers crevassing stops and plastic flow commences has been found by seismic methods at Crillon Lake, Alaska, to be 30 m.[297] The minimum load necessary to produce continuous creep or extrusion flow may be of the order of 45 m of ice in depth as estimated by Demorest or even less,[298] depending probably, among other factors, on grain size.

Translation. The physical properties of ice depend upon the orientation of its grains[299] as J. Brewster[300] first demonstrated. Pressure applied perpendicularly to plates or laminae (which are parallel with the basal plane (0001) and lattice slippage) induces elastic bending[301] with undulatory extinction, but when applied parallel with the laminae gives rise to gliding or "translation" with no optical change or distortion of the lattice. This is due to the readiness with which the sheets of hexagons (see p. 109), although puckered, glide over one another. Since grains in glaciers are promiscuously orientated, some laminae are bent and exhibit undulose extinction and others are translated without distortion.[302]

The crushing strength, normal to the principal axis, is only one-third of that when pressure is applied along the axis.[303] Thus translation, which is known for many other substances, such as iron, copper, gold and silver and causes ductility in metals, has been experimentally proved on glacier-ice

after a certain minimum pressure has been exerted.[304] The differential flow of glacier-ice is to some extent analogous to the influence of rolling on a multi-crystalline sheet of a metal of hexagonal symmetry such as magnesium.[305] Slip within the grains may also possibly take place along planes other than the basal plane,[306] including internal fracture surfaces.[307]

This translation has been recognised as the cause of ice-flow.[308] It has been thought to take place at a depth of 17 m (O. Mügge) where there is the requisite minimum pressure or in the zone of shear in the basal parts or lower ends where granular interlocking is complete.[309] Yet experiments[310] compelling bundles of ice-crystals, commonly orientated, to move either along glide planes or between crystals, show that lamellar gliding is not important and that ice is deformed by adjustment along the contact surfaces of adjacent planes.

Regelation. M. Faraday[311] demonstrated that two pieces of ice in contact at 0°C froze together without increase of pressure. This "regelation", as Tyndall and Huxley named it,[312] is vital in a glacier's economy; by its means snow passes into névé, crevasses heal, contiguous glaciers unite, and ice-avalanches fuse into reconstructed glaciers. While the path along which the wire in Tyndall's well-known experiment cuts the grains is traceable by a band of included bubbles,[313] the ice loses none of its strength and its optical orientation is undisturbed, so that in reforming above the wire it is controlled by the orientation of the bordering crystals.

Although regelation, which is facilitated when crystals are similarly orientated,[314] has been regarded as the cause of ice-motion, there is a concensus of opinion that it is only a concomitant or secondary factor arising from flow.

While pressure is not essential for regelation, as affirmed by Faraday[315] and Tyndall,[316] proved experimentally by Emden[317] and explained by L. Pfaundler,[318] regelation takes place more readily if pressure is varied.[319] N. L. S. Carnot demonstrated that pressure lowers the freezing point of water and C. Hutton[320] concluded from experiments that, by checking expansion, it prevented freezing, whatever the degree of cold. J. Thomson[321] showed as a deduction from the mechanical theory of heat that by partially liquefying the ice it aided regelation. This was proved simultaneously and theoretically from the second principle of the mechanical theory of heat by R. Clausius[322] and later by A. Mousson[323] and W. Thomson[324] who ascribed the change of form to incessant liquefaction of ice at places of intense pressure and the prompt refreezing of the waters thus set free in other and lower parts when the pressure was relieved, a view supported by H. v. Helmholtz.[325] This liquefaction of ice by pressure, witnessed on modern glaciers,[326] is made possible by the very open structure of ice (which incidentally also accounts for the abnormal expansion, instead of contraction, of water on freezing).

J. Thomson[327] from theoretical considerations predicted that as ice contracts in volume on fusing, its freezing point would be lowered 0·0075°C for every atmosphere (later figures[328] range from 0·00722 to 0·00753). This "Thomson effect" was afterwards verified[329] on the Arolla Glacier and Hintereisferner: at the bottom of the Greenland ice-sheet the melt temperature has been calculated to be −1·2°C.[330] Yet the melting point does not vary uniformly with pressure. This was proved by Mousson[331] and afterwards more fully by G. Tammann[332] who showed the connexion between the

pressure necessary for melting ice down to a temperature of $-22°C$, below which pressure cannot liquefy ice (see p. 44). He found the lowering between $0°C$ and $-2\cdot5°C$ was 0·0076, between $-2\cdot5°C$ and $-5\cdot0°C$ was 0·0093. These figures, however, only hold if the melt-waters cannot escape; for when they do the lowering is 0·0088 between $0°C$ and $-2\cdot2°C$. Non-uniform pressures, such as prevail in glaciers,[333] lower the melting point at a much higher rate, even ten times as great and up to $0\cdot09°C$.[334] Experiments[335] prove that as the temperature falls the pressure required to produce melting rises rapidly until it exceeds the crushing strength. This pressure-liquefaction theory of ice-flow, especially in the basal layers, has often been advocated.[336]

Experiments demonstrate that welding occurs at temperatures sufficiently low to preclude the possibility of uniform pressure melting and regelation.[337] Pieces of ice in north Greenland freeze together at $-15°C$ to $-50°C$[338] and J. Vallot observed coarse-grained névé in 1897 in tunnels near the summit of Mont Blanc where the annual temperature was $-16\cdot5°C$ and the pressure only one atmosphere.

That regelation is not a necessary condition of flow has been shown for Spitsbergen where the ground under the glaciers is frozen to a considerable depth and the temperature is always below $0°C$[339] (cf. p. 108). Hess[340] thought the plasticity was "dry" as the coefficient of internal friction is practically the same whether the temperature is at or below melting point. That there can be translation without regelation is shown by the occurrence of translation in substances which lack this property.[341] Wright and Priestley,[342] in their thermo-dynamic theory of plasticity and ice-motion, think pressure does not cause melting but increases the number of molecules which tend to diffuse to parts less favoured. Crystal boundaries are the seats of molecules of energy-content above the average and include mobile molecules about to escape from one crystal into an adjacent one. If the surface-energy of the crystal differs for different faces, the number of mobile molecules able to pass the boundary will depend on the crystal's orientation with respect to adjacent crystals. Big crystals have smaller mean energy per molecule so that they grow at the expense of small ones. Rise of temperature makes the mobile molecules more numerous and leads to increased growth. E. K. Pyler,[343] however, sought to show experimentally that it is unnecessary to assume two kinds of ice-molecules.

Granular theory. The granular theory, consistently advocated by Chamberlin,[344] had for a time the support of the majority of modern glacialists[345] who tended to abandon the viscous theory largely under the influence of his work. The theory supposes that glacier-grains are the mechanical units of motion and, like a mass of shot, slide over one another along the surfaces of least cohesion without actual fracture, as envisaged, for example, in G. Quincke's "foam-cell" hypothesis (see p. 113). Ice is a rigid if weak crystalline rock of the purest and simplest kind. Displacement is permitted by the momentary liquefaction (actual melting may not take place, merely slow idiomolecular transfer) of minute parts of the granules at points of contact and compression, and by transferring the water to adjacent parts not in contact where it solidifies. The ice as a whole remains rigid, but behaves under deforming stresses according to the same mechanical principles as other rocks and moves by recrystallisation, granulation, constant readjust-

ment of granulated particles, gliding in the grains themselves and shearing. In the process, the grains lose or gain in size and alter their relative positions. By the summation of their slight adjustments, the mass flows along lines of least resistance. The impurity and salinity of the inter-granular film, which pressure alternately thins and thickens, may play an important role by lowering the melting point.[346]

Whether this movement is a *vera causa* depends obviously upon the degree of interlocking and the molecular cohesion at the intergranular surfaces. In the névé, the grains are able to rotate and roll over each other either as individuals or in clusters to supply the chief cause of flow. The irregular interlocking in the tongue renders such flow here less easy. Intergranular movement may, indeed, be prevented[347] since the intergranular film is not thick enough to permit the degree of articulation required for continuous flow. Moreover, the coefficient of internal friction of the ice is that of the individual grain and the plasticity lies within the grain.[348] Nevertheless, intergranular movement does seemingly take place, where interlocking is not intimate. It is facilitated in the snout by abundant lubricating water and in the deepest layers by pressure which by transferring material weakens the granular attachment and permits some slipping and rotation of the grains: striations on granules have been attributed to viscous shear.[349]

The motion of each individual grain requisite to give the known flow is very slight. In a glacier 6 miles (*c.* 9·5 km) long with a daily flow of 3 ft (*c.* 90 cm) a grain would have to move the length of its own diameter with reference to its neighbour once only in 30 years.[350]

Glide planes. It has frequently been averred,[351] especially by Philipp,[352] that motion is by continuous and intermittent slips along multitudinous glide planes, 0·5–2 m apart, arranged trough-shaped, particularly near the base and margin, i.e. between the more slowly and more rapidly moving ice where the elastic limit has been passed.[353] Like any other rock, a glacier is rigid up to a certain point beyond which it yields under stress. Such shearing has been shown to exist[354] and is compatible with pollen analyses of adjacent layers of ice in Alpine glaciers.[355] Glide planes themselves have often been observed,[356] e.g. in the Alps (here they are in fact rarely seen because the glaciers are thin and strongly crevassed and have abundant debris[357]), Karakorams and Pamirs, Scandinavia and arctic regions (calculations show that gliding forms not a little fraction of the total flow in Greenland), in South America, and in the Antarctic. The differential movement has been measured on the Victoria Glacier, Canadian Rockies.[358]

The planes, which often tend to coincide with the banding in glaciers and with one of the slip-planes,[359] are often clearly discernible. The upper layers project above definite planes, as in the ends of Greenland glaciers or the sides of ablation valleys in the Karakoram Mountains[360] where they mark merely the trace of the plane of one glacier resting on another. Ground-moraine is extruded along them[361] though this, which may facilitate ice-avalanching,[362] has been ascribed to a difference of ablation as between clear and silt-laden layers. The planes sweep across the surface of glaciers parallel with the margin. They are accompanied[363] by faulting and drag phenomena and by structures resembling ordinary fault breccia; by twisting and distortion, readily recognisable if the ice-layers are associated with bands of debris; by smoothed surfaces resembling slickenside; by fluting on the interpolated

laminae of debris; by flat boulders, faced parallel with the planes of dirt layers; by long drawn out bubbles of air in the ice; and by major planes of slipping which are veritable thrusts of considerable displacement—in the Karakorams, these are apparently not uncommon along the planes of contact of glacier components.[364] Striated pavements and englacial scratching of pebbles may have been done in this way.

Glide planes are also found in névés, particularly on steep inclines.[365] Glacier movement, e.g. in a cirque basin, below an ice-fall or at the snout, may take place in part by rotational slipping along glide planes[366] which in some cases pass from stratification planes. Basal slip, more or less following the rock-bed, may in the case of small glaciers cause the whole glacier to rotate like an unstable clay slope, the bergschrund being a vertical tension crack. The weight of the thicker mass above the firnline (see p. 27) and the melting at the lower end cause the firn to flow out and beneath, so that the maximum flow is in depth and in winter.[367] Experiments show that creep occurs at temperatures down to $-20°C$ even under low loads, so that shallow glaciers like the Claridenfirn and much more so ice-sheets would sag under their own weight, so that extrusive flow would occur.[368] Nevertheless, it is quite unwarranted to assume with Philipp[369] that stratification is due to gliding, especially since no slip has been detected between layers in a névé's upper part.[370]

Movement along glide planes is facilitated[371] by circulating waters and saline matter, and between different layers by dirt which depresses the melting point and prevents the interlocking of grains. These apparently do not cross the planes[372] unless perhaps in the older banding.

It has been regarded as axiomatic since the early labours of Agassiz[373] and Forbes[374] that ice-motion is continuous apart from any irregularities that may attend the opening and closing of crevasses. Observations on glaciers, the appearance of the painted stones below the great ice-fall on the Rhône Glacier in great curves just as might have been predicted,[375] and the nature of striations—all these seem confirmative. Yet it has been believed with de Saussure that the motion was jerkwise.[376] This has been suggested by Greenland observations[377] and by careful measurements on a number of glaciers,[378] though other observations and P. L. Mercanton's *cryocinémètre* (a piece of apparatus devised to magnify movements) seem to disprove it.[379] Further accurate, preferably continuous and self-recording observations are necessary to solve the problem.

On the Hintereisferner and certain other glaciers,[380] the flow consisted of jerks or pulsations with increased velocity and intermediate pauses of small velocities due to "pressure waves"—not to be confused with the *Schwellungswellen* produced by changes in velocity consequent upon increased loading of the firn which have a velocity 20–150 times that of the ice (see p. 155). These "jumps" seem to show that ice yields to stress and that this cannot greatly accumulate without causing displacement. Stress develops locally until it demands adjustment, either by sudden slip along fracture planes or by short, rapid yielding.

The abnormally high rates of movement of the sides of the Rakhiot Glacier on Nanga Parbat, in Jostedalsbrae in Norway, in Karajak and other glaciers of Greenland, and in Tindbrae and Kings Fjord Glacier, Spitsbergen, have shown that on some fast-flowing glaciers the ice flows between two narrow deformation zones, a few tens of metres wide, by a block movement,[381] quite

different from the normal continuous flow in slow-moving glaciers—examples were found in the Alps in earlier times, e.g. in the Vernagtferner which in 1845 had a speed of 11 m a day. The consequent deformations bring the ice up to the critical point of stress-strain established by E. Orowan[382] theoretically and experimentally. The ice is broken up into *Schollen* of various sizes which move as units like solid bodies influenced by irregularities of the bed. Extremely broken surfaces and numerous gaping crevasses, ice-pinnacles and séracs are the outward signs of this *Block-Schollen* type. One part of a glacier may move by *Block-Schollen* flow, while other parts have a pure stream flow.

That glaciers move along glide planes is denied.[383] These it is said are local and rare and restricted to the extremities where the ice is frozen to the ground in winter, and do not occur near the base where the pressure is high and the ground unfrozen. Moreover, cohesion, as laboratory experiments show (their applicability is controverted[384]), is no less along these planes than in other directions. Nevertheless, such fracture or differential shear does undoubtedly exist,[385] especially in the more rigid arctic glaciers and in the brittle outer crust and where the normal flow is impeded. It is apt to occur along the curved sole where englacial debris reduces plasticity and where the rate of flow, both horizontally and vertically, changes most quickly (see p. 102), and where the active ice behind presses against the thin and rigid or inert snout and either overrides or pushes it forward bodily. Whether it is a prime cause and works within the main body is less sure: here the flow is seemingly plastic[386] (pl. VIB, facing p. 113).

In fine, ice moves by a combination of processes, partly by sliding upon its bed and principally by plasticity arising from intergranular and intermolecular movement and by movement between clusters of crystals, and by intermittent shear along glide planes, all facilitated by regelation. Each kind of movement probably exists in different parts of one and the same glacier[387] and varies with temperature, pressure, stress and other conditions.

Finsterwalder's theory. Finsterwalder's mathematical or geometrical theory[388] made a great advance in our knowledge of the glacier. It is free from the questionable or arbitrary assumptions concerning the physical properties of glacier-ice which his precursors[389] along these lines had made, though it does not take inner ablation into acount. In endeavouring to apply a geometrical treatment to the question of ice-flow he started with an ideal glacier, i.e. one in which velocity is constant, continuous and independent of time and all the contour lines on the entire glacier remain in the same position. Streaming continues unchanged in direction and strength, neighbouring particles remain in steady contact, free from eddying, and there is no sudden jump between contiguous layers. Each particle from the névé, after pursuing a definite course in the interior, emerges at a definite point on the tongue. Finsterwalder joined all such points of entrance and descent in the névé with corresponding points of emergence in the tongue by lines of flow—others had already begun their study.[390] Lines beginning at the firn's edge run along the base to emerge at the margin and others starting out of the firn pass through the glacier to appear at corresponding points on the tongue. Flow-lines, therefore, emerge all over the tongue, including its sides and snout. Streamlines or the paths of particles pass downwards in the névé (also in the "regeneration region" of glaciers without firn[391])—the deepest firn crevasses

are free from debris—and pass upwards in the tongue at an angle depending upon the rates of ablation and of flow—earlier observers had already noted the vertical component.[392] The paths of bodies on the glacier surface (*Bewegungslinien*), mapped with great accuracy on the Hintereisferner and Rhône glacier[393] but as yet unexplored on the ice-sheets, do not coincide with such

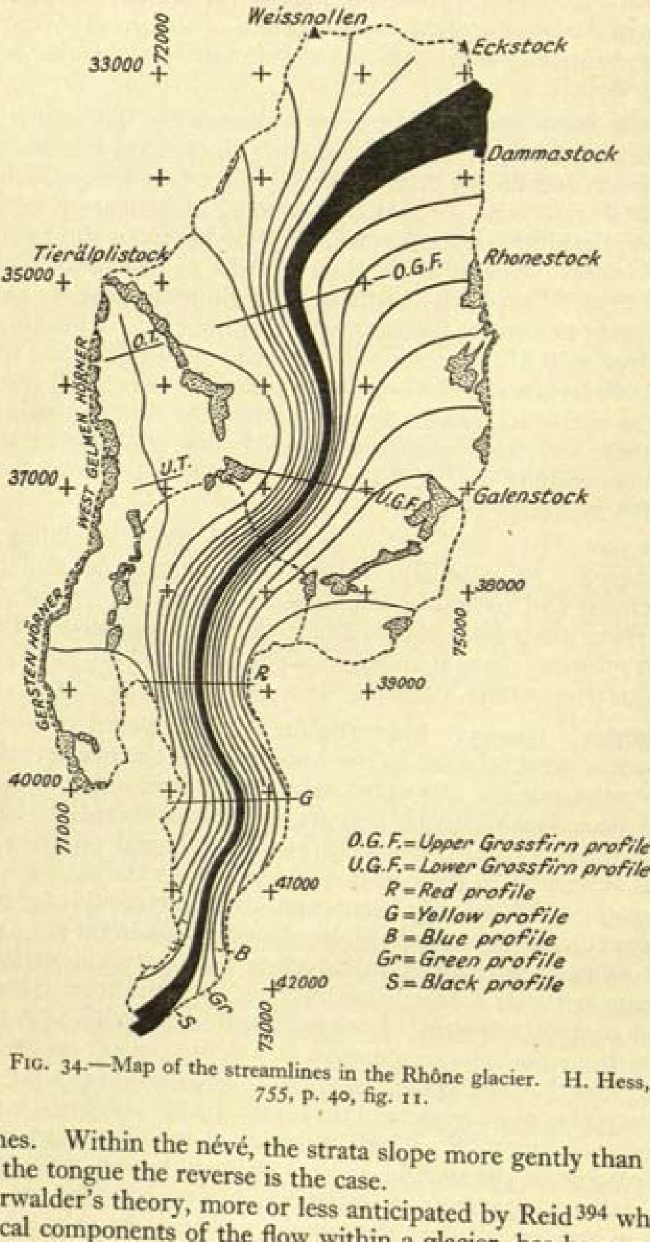


FIG. 34.—Map of the streamlines in the Rhône glacier. H. Hess, *755*, p. 40, fig. 11.

streamlines. Within the névé, the strata slope more gently than the streamlines: in the tongue the reverse is the case.

Finsterwalder's theory, more or less anticipated by Reid[394] who measured the vertical components of the flow within a glacier, has been confirmed by bores and by measurements of flow on the Hintereisferner,[395] by stone-lines on the Mer de Glace, Vernagtferner and Rhône Glacier (fig. 34), and by

pollen analysis (see p. 50). Thus the trajectories in the Rhône ice-fall are not crossed: the chaos is only at the surface.[396]

Finsterwalder's theory explains satisfactorily many well-known facts: the carrying down into the ice of detritus which falls on the firn and the remaining upon the surface of the debris below the reservoir; the passage along a glacier's bed of material which falls upon the margin of the firn; the spreading of moraines near the snout; the movements of the ice and the influence of abrasion; the upturned layers at the snout; the formation of dirt-bands; the spoon-shape of banding; and the waves in an advancing glacier. It is however descriptive and not a quantitative theory like those later formulated[397] on a hydrodynamic basis, though these are irreconcilable with the fact that the viscosity is not constant but decreases as a high power of the applied shear stress.[398] Somigliana's theory (see page 38) applies only in stationary glaciers with regular bed, constant surface gradient and constant velocity, and if the velocity at the margin is nil.

Flow in Pleistocene ice-sheets. As observed already, each theory of glacier-flow has a grain of truth: it is valid in part and in certain circumstances. The only common feature is the motive power of gravity which T. Vidalin in 1695 first invoked for ice-flow.

The figures of flow of the present diminutive Alpine glaciers are inapplicable to the vast masses of the Pleistocene. Nor may those of other regions be used without modification; Iceland because of its vulcanicity, Alaska because of its climatic conditions, Greenland because of its peripheral ring of mountains and glaciers squeezed at great pressure through gaps, and the Antarctic on account of its starvation by extreme cold. The lower latitude and higher temperature in ground and ice, the higher levels of the valley-floors before the ice lowered them, and the much deeper ice (see p. 40) and concomitant higher plasticity make it highly probable that the Pleistocene ice-sheets flowed more rapidly.[399] J. Vallot,[400] from formulae linking velocity with thickness, computed the daily velocity of the Pleistocene Rhône Glacier (1000 m thick) at 4·5 m and that of the Mer de Glace (500 m thick) at 2·9 m. O. Ampferer's estimate for the Pleistocene Inn Glacier was 3 m[401] though A. Penck[402] thought the Swiss glaciers generally had velocities comparable with present-day Greenland (see p. 104): he instanced the annual velocity of the Rhône Glacier at 600 m, of the Isar at 160–320 m, and of the Rienz at 100 m. Similar calculations[403] have been made for the Pleistocene Rhine and Rhône glaciers and for the Muksu Glacier, north-west Pamirs. J. D. Dana,[404] however, imagined the ice-sheet moved daily or weekly *c.* 30·5 cm and W. Upham[405] that its base had a maximum annual velocity of 15–30 m.

The flow was, of course, neither constant nor uniform. It was probably quicker over western Norway than over the Baltic or the plains of Russia and Siberia and was faster on the southern margins, especially in summer. Exigencies, such as inclination and inequalities of the ground and varying depth, led to differences and to a more rapid flow along the valleys and depressions—erratics and drifts were carried farther in the basins than on the ridges.[406] It varied too with the nearness of free ice-discharge. Thus the sea affected the flow and reach of the ice-sheets. If the ice deployed in the sea, it was floated off or melted and had a more rapid flow imparted to it since the more slowly moving lower part was removed to allow the quicker ice to

flow at its own rate[407] in much the same way as the speed of the Alpine glacier above an ice-fall is increased after the ice has broken away. In the interiors of the ice-sheets, with their very low surface-gradients and basined floors due to isostasy, the flow was extraordinarily slow: the velocity across any section was approximately proportional to the radius of the section.[408]

The problem of the flow of the Pleistocene ice-sheets bristles with such difficulties that some geologists[409] have even denied the very existence of these sheets in Europe and North America. Experiments on the modulus of cohesion led R. D. Oldham[410] to believe that ice could not move *en masse* over a plain to any great distance because its lower layers would stagnate. It has recently been suggested that the ice was passive: the apparent southerly flow over most of north-west Europe was due to a northerly movement of the continent under the ice.

All hypotheses agree, as noticed already, in believing that the weight of the ice is the prime cause of motion in modern glaciers. In steeper valley glaciers and the outlet glaciers of Greenland movement is by gravity flow. The efficiency of gravity in the sliding hypothesis is due mainly to the state of the glacier's sole, in other hypotheses to conditions within the glacier itself. In the case of the Pleistocene *mer-de-glace*, the slope of the ground exercised little influence, save during the early and late phases and often in valleys near the ice-edge, as in the "basins of exudation" in Greenland to-day or the Norwegian Channel and Hudson Strait of the Pleistocene.[411] Any slight initial elevation was counterbalanced by the resistance offered by the varied and complex relief over which the ice moved. In these cases, movement may be by extrusion flow,[412] either obstructed or free, as it is in expanded-foot and piedmont glaciers. Obstructed extrusion flow occurs where there is a physical obstacle or a thinning of the ice, e.g. towards the margin and in the dissipation region, and gravity flow which is a drainage-controlled flow as in a river, e.g. in valley glaciers or outlet glaciers of an ice-sheet. The greatest surface velocity is towards the margin and since this requires a steeper surface slope, the profile becomes more and more convex.

The theory of extrusion flow was arrived at by M. Demorest from observations on glaciated terrain in North America and from observations on the ice of Greenland, and by A. Streiff-Becker from calculations of the combined effects of ablation, settling and flow on the one hand and the annual accumulation on the other on the Claridenfirn of the Alps which gave a speed of transport of snows greater than the measured flow at the surface. Nevertheless the theory has been criticised[413] because, first, the strength of the ice which is subjected to plastic flow is slightly greater than that of the ice at the surface since its smaller crystals give added strength to the mass; and, secondly, because the bottom flow is not greater than the surface flow as displayed by the behaviour of the 1000-ft deep bore in the Malaspina Glacier (see p. 37). The maximum flow, it is said, is not at the base but at the level of the outlets from the interior basin.

Highland centres in Pleistocene times were not always effective dynamic centres of propulsion as Close[414] was one of the first to realise. The ice flowed over vast plains, as in west Canada and north Germany and Russia, against the slope of the ground, under conditions which may have no modern counterpart—in North America the height at the Keewatin centre was below 500 ft (*c.* 150 m), at the Labrador centre 500–1500 ft (*c.* 150–450 m).[415] The instances, purely local and uncommon, of uphill flow over rising ground,[416]

due to the longitudinal force, are not helpful because the rise is small and the elevation and thrust from behind is great. Even the till's lubricating action sometimes invoked[417] is of secondary import.

The *modus operandi* may be plasticity resulting from one or all of the causes previously set out (see p. 113). But what impels the mass forward? This question has been answered in two ways; the first supposes that gravity was the essential force and that the northern lands which acted as ice-centres were higher though subsequently they were lowered by subsidence or ice-erosion[418]; the second postulates that the ice was extremely thick and flowed by its own weight, that is hydrostatically or by extrusion.

A great increment in height at the ice-centres as the cause of glaciation and ice-flow was postulated for the Alps,[419] north Europe,[420] British Isles,[421] North America[422] (North America and north Europe were glacierised from an Atlantean Continent, 17,000 ft or 5180 m high[423]), Siberia,[424] New Zealand[425] and east Africa,[426] and as part cause for the greater Antarctic glaciation.[427] From the time when Lyell and Dana began to give it their consistent support up to 1875 this postulate had a majority of adherents. The amount was rarely stated but ranged up to 5000 m for Scandinavia, 1500 m for the Alps,[428] and 5000–7000 ft (1525–2135 m) for North America[429]: the uplift was sufficient, it has been said,[430] to divert the north German rivers into Bohemia and Moravia and enable the melt-waters to excavate the canyons of Sächsische Schweiz.

It was maintained that the uplift ensured the necessary low temperature and snowfall[431]; reduced the thickness of ice otherwise necessary; and avoided a flow against the slope of the land.[432] Yet the arguments against the upheaval as a general cause of ice-flow and of glaciation are overwhelming: we shall consider them in later chapters (see chs. XLIV, LI). All we need say at this place is that the lands, isostatically depressed, were almost certainly lower than now; and that the required inclination could not have been given by any *probable* elevation.

The second method of imparting the tremendous impelling power or *vis a tergo*, the importance of which was early recognised,[433] presumes that flow depends less on land-levels than on the depth and pressure of vast snow accumulations piled up in the interior[434]; these cause compressive stress and lead to outward movement over horizontal or even reversed surfaces. The Hintereisferner[435] shows that gradient is subordinate to cross-section and thickness in influencing the flow. Large glaciers move more rapidly than relatively steep tributary glaciers with less depth and cross-section.[436] Ice, if thick enough, may flow uphill; for the determinant is the gradient of the ice-surface in relation to the reversed bottom slope, the "specific angle" as it has been termed.[437] In ice-sheets, there is a downward movement in the central area that passes into a horizontal and outward movement in the peripheral zone.

But some glacialists[438] consider that even thicknesses of several thousands of feet, unless the basal layers be very plastic, cannot give the required propulsion. For example, the North American ice-sheet, as stated by W. H. Hobbs, may be represented in cross-section by a line 6 in. (15·2 cm) long and 0·01 in. (0·254 mm) thick[439] and the gradient of the ice-surface from Scandinavia to the German Mittelgebirge may have been only 0° 3′[440] (fig. 35). Yet flow takes place in Greenland with slopes of only 0° 1′.[441]

To overcome this difficulty, which is but enhanced by "basining" or the

effects of isostasy which depressed the crust in the centre (the floor of the Baltic may have been depressed 500–800 m, cf. fig. 35) and (possibly) raised a marginal bulge (see ch. XLVI), as well as the natural reluctance to conceive an ice-sheet so thick that it could control the flow from its centre, it has been suggested that the ice did not move outwards throughout its mass but was immobile in the centre and between the hills. Flow was restricted to the margin,[442] estimated at 30–40 or 100–120 miles (*c.* 48–64 or 160–192 km), and, as the pronounced marginal gradients indicate (see p. 42), was induced by the snow which fell upon it[443] or the "centrifugal broom" swept to it.[444] After its early stages, the ice-sheet grew and spread from the marginal snows rather from the outflow from central ice-fields.[445] The intramarginal belt,

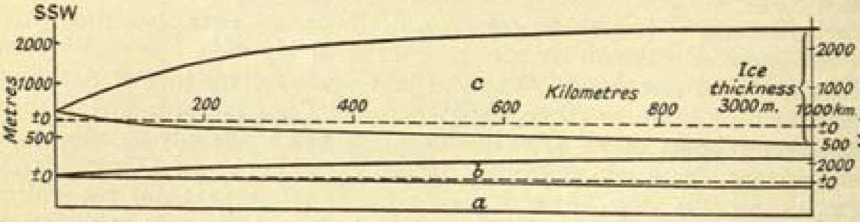


FIG. 35.—Section through the Scandinavian ice-sheet at Elster maximum from the Åland islands to the Sudetes. *a* = to natural scale; *b* = vertical scale 10 times exaggerated; *c* = vertical scale 50 times exaggerated. P. Woldstedt, *P. M.* 96, 1952, p. 268, fig. 1.

especially that on the windward side, and not the geographical centre controlled the movement. By thickening, the ice-sheet caused not only an outward flow but an inward flow until the surface became nearly horizontal throughout its entire extent.[446] Several facts are urged in its favour. The snowfall on modern ice-sheets has an analogous distribution (see p. 665); the centres of the Greenland and Antarctic ice-sheets are without crevasses (see p. 47); and boulder-clay is mainly local (see p. 379). Nevertheless, this view is inconsistent with the laws of mechanics as applied to plastic solids, and with the currents of ice which converge upon the "dimples" and come from far back in the interior. Pressure ridges as far as about the South Pole prove that the ice is moving on the Antarctic plateau. Seemingly contradictory too are the far-travelled erratics,[447] including those which have moved 1000 miles (*c.* 1600 km) from James Bay to North Dakota, Minnesota and Lake Superior[448] (cf. enumeration of other North American instances[449]) or from the Adirondacks into the oldest drifts[450] of Pennsylvania and Kentucky. In Europe, rocks from Elvdalen in Sweden, from Umptek in the Kola Peninsula, and from the Dalarne and Åland Islands have been carried almost 2000 km.[451] This objection, which implies that the precipitation was no heavier at the margins than at the centre,[452] is not insuperable since purely marginal flow, with an expanding ice-sheet, might convey central boulders in interrupted journey to the periphery.

There is undoubtedly in the interior of the ice-sheet a thick unfrozen firn susceptible of little plastic deformation. This however rests on the basal layers which are plastic and flow outwards (as Rink[453] and E. v. Drygalski[454] maintained) to feed the outlet glaciers and the marginal part of the ice-sheet.[455] The central firn area, therefore, except in the very centre where the outward flow will be negligible and approximate to the conditions postulated by Hobbs, is probably subsiding very slowly to compensate in part for this

outward peripheral movement, though its surface may be maintained at roughly the same elevation by the heavy accumulation.

The hypothesis[456] that the ice east of the Scandinavian ice-divide was a *braeplatte*, passive and thin, pushed forward by new ice on the parting where precipitation was heavy has little to commend it.

Flow in the Pleistocene ice-sheets was facilitated by subglacial waters[457] (cf. p. 111) and made possible by the growth of very deep ice at the centres which raised the temperature and plasticity of the basal layers. The temperature throughout much of the mass was not far removed from fusion point. The molecules were highly energised by heat so that their cohesive force was almost neutralised and the flow continued indefinitely without arrest.

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Chapter VI

METEOROLOGICAL AND GLACIER OSCILLATIONS

1. *Meteorological Periodicities*

Brückner and other cycles. Variations of the various meteorological elements, other than those of the day and season, have been held to exist since the days of Sir Francis Bacon[1] (1625). F. W. Ehrenheim[2] in 1824 gave the first analysis of historical reports bearing upon the subject and about fifty years later (1873) W. Köppen[3] correlated temperatures with sunspot cycles (in 1844 Gautier had sought such a correlation for the temperatures of Geneva). Search for periodicities in modern meteorological records has enjoyed quite a boom during the present century among meteorologists, climatologists, geographers and others, and something like 150 cycles have now been found. J. Hann[4] has summarised, with literature, the methods and values of calculations and N. Shaw[5] has listed the amplitudes and lengths of the empirical "cycles" derived from an examination of long series of observations by inspection or arithmetical manipulation. Unfortunately, even the oldest meteorological records cover only about 250 years (Paris 1664, Breslau (Wroclaw) 1692, Berlin 1700, Uppsala 1722, Lund 1723, Padua 1725, Leningrad 1725 and New Haven 1780).

Brückner[6] investigated some of these variations for Europe and for certain extra-European localities. He thought they conformed to a cycle, since known as Brückner's cycle,[7] of approximately 35·5 years (or 34·8±0·7), the hemicycle varying between 20 and 5·5 years. His data concerned air pressure, temperature and precipitation; the frequency of severe winters; the levels of lakes and seas, e.g. Baltic, Caspian and Black Sea, and the English Channel at Brest, Cherbourg and Le Havre; the disappearance and reappearance of lakes; the break up of ice on rivers, as in Russia; vine harvests and wheat prices in west Europe and the oscillations of climate and immigration into the United States. He found the following wet and dry periods:

Wet Periods	*Dry Periods*
1691–1715 (*c.* 1705)	1716–1735 (*c.* 1720)
1736–1755 (*c.* 1740)	1756–1770 (*c.* 1760)
1771–1780 (*c.* 1775)	1781–1805 (*c.* 1790)
1806–1825 (*c.* 1815)	1826–1840 (*c.* 1830)
1841–1855 (*c.* 1850)	1856–1870 (*c.* 1860)
1871–1885 (*c.* 1880)	

Since Brückner's paper of 1890 focused attention upon the question, his cycle has been widely recognised in meteorological and related matters in all quarters of the earth. In illustration we may mention epidemics[8] and Russian famines[9]; temperatures[10] at Oxford and in central Siberia; pressures in all quarters of the globe[11]; wind velocities[12] at Greenwich and in Austria; precipitation,[13] e.g. on the northern margin of the Alps, in Austria and Hungary, in Jugoslavia, over Europe and in the United States; levels of lakes[14] in Greece, Armenia, Caucasus, Scandinavia, the Baltic, Starnberger See, Victoria and Albert Nyanza, Lake Nyassa and the Great Salt Lake of

Utah; strong floods[15] in the south-west Baltic, Nile and Moldau; mud-slides[16] in Switzerland; dune-formation[17] along the south Baltic coast; vine harvests and crops[18] in Sussex and Ohio; tree rings[19] in Switzerland and western U.S.A.; freezing over and break up of the ice[20] on the upper Danube, on Finnish rivers and on certain Alpine lakes; and drift-ice[21] near Iceland (due to the varying strength of the North Atlantic pressure distribution and circulation)—in ice-poor years east winds weaken the East Greenland Current and press it and the ice against Greenland while in ice-rich years stronger north-east winds produce a greater intensity and distribution of the stream and a greater transport of ice into more southerly latitudes. The Humboldt Current is deflected westwards at intervals of about 34 years[22] and Weddell may have penetrated the Weddell Sea at a Brückner's maximum[23] (though Weddell's voyage into this sea has been regarded as fictitious[24]). The cycle is thought to have a solar origin[25] and to have extended backwards into the postglacial dunes of the Pomeranian coast[26] and into Pleistocene time in Co. Kerry (see p. 144) or Jurassic time in Swabia.[27]

Similar cycles have been thought to exist in the rainfall[28] of Lombardy, the precipitation and temperature[29] of Brussels and Russia and the Etsch region; the levels[30] of Lake Constance and of Lake Ngami in the Kalahari; the sea[31] along the north French littoral; the bottom muds[32] of the Sea of Aral and other lakes; Alpine avalanches[33]; and the limits of polar ice[34] which coincide with the periodicity of the north Siberian river-ice.[35] Graphs of the approximate changes in the Caspian Sea since 400 B.C. agree with those from the whole of west and central Asia.[36]

The range of temperature within a cycle which Brückner[37] gave as 1°C is probably less: S. Newcomb[38] found it to be only 0·26°C in the tropics and middle latitudes and C. Nordmann[39] obtained 0·33°C. W. Köppen[40] calculated its amplitude in the temperate zone to be 0·43°C and in the tropics 0·59°C and thought it was scarcely perceptible in polar zones though F. Dilger[41] believed it became bigger with latitude.

The rainfall oscillation is much greater and increases with continentality[42]; in Europe it is 14–20% of the mean annual fall, in continental Asia 36% and exceptionally 100%.[43]

Brückner's cycle has been denied,[44] e.g. for the severe and mild winters of Berlin, for the temperatures of Sweden and the weather in west Europe, French Alps and Moscow, for the level of the Caspian Sea and for sea-level off the German coasts. It has been found inconclusive[45] for the climate of U.S.A., European rainfall or Finnish floods or is said to occur at other than Brückner's dates,[46] as in the severe winters of European Russia. More generally, the cycles are of unequal length; wet and dry periods do not coincide with cold and warm periods[47] and, contrary to Brückner's opinion,[48] are neither universal nor contemporaneous over the earth[49]—dry periods in lower latitudes correspond to wetter periods in higher latitudes[50] and Europe and North America are also in opposite phase[51] (see p. 138). Moreover, the periods are probably not pure but consist of shorter ones of different lengths superimposed one upon the other.[52]

Shorter fluctuations have indeed often been described.[53] Such are, expressed in years, the 1·7 and 2·11 (with 4·7, 9·5 and 51·7) for British rainfall[54]; 2 for temperatures at Oslo and Bergen[55] (with 4, 8, 11 and Brückner's cycle); 2·2 for Greenwich temperatures[56]; 2·75[57] for the tropics and levels of Vänern and the Rhine (at Basle); and the rainfall of Australia, the

tropics and the North Atlantic; 3 for pressures in the East Indies[58]; 3·2 for precipitation[59]; 3·32[60] for tree-growth in Java; *c.* 3·5[61] for pressure and atmospheric circulation; 3·8[62] for pressures in all parts of the globe; 4–5 for the atmospheric circulation and dry years in England,[63] for recurrent cold winters,[64] as in Sweden and England; for ice-conditions[65] in the south-west Baltic and Kattegat, Iceland, the Greenland seas, and in most of the Arctic and north Asia, corresponding to the periodicity in atmospheric circulation A. Defant[66] described; and 5·25 for rainfall, temperature and pressure in the North Atlantic region.[67]

The 5·5 or "Hellmann relation", which is presumably the 11-year cycle (see below) with two maxima usually unequal, has been detected[68] in north Germany and in tree-growth in Europe and Arizona; 6 years 1 month in the rainfall of the Pacific coast of North America[69]; 7 years in various terrestrial weather phenomena[70]; 8 in Swiss pressures and temperatures, in crop yields and rainfall in U.S.A. and the deep-sea currents in Norwegian fjords[71]; 6–18 in the cold winters of Russia[72]; 10 in the ice-formation in Barents Sea[73]; 10–14 in the levels of Lake Michigan[74] and 11–13 in the Great Lakes[75]; 8 and 16 in Vänern[76]; 16 in tree rings of Tauern and Viennese temperatures[77]; 19 (due to the effect of the moon) in pressures in Australia and South Africa[78]; and in the five continuing, regular periodicities of the sun's radiation[79] (8 months, 11, 25, 45 and 68 years).

Trees, especially pines (*Pinus ponderosa*, *P. edulis*, *P. monophylla*), fir (*Pseudotsuga douglasii*) and sequoias (*Sequoia washingtoniana*) in central California, which grew between latitudes 36° and 39° N. on the outer slopes of the Sierra Nevadas at 5000–7000 ft (1525–2135 m), have an immensely long life and annular rings dating back 3233 years. Since the width of the rings is controlled[80] by illumination, ground-water supply, and principally by temperature (as in the damp climate of west Europe) or precipitation (as in the dry areas of Arizona), or by the interaction of both, they afford a useful means of determining the climatic variations of the past. By this dendrochronological method,[81] the biological counterpart of varve chronology, applied to more than 500 trees, and by proceeding from living trees to historic or prehistoric timbers, A. E. Douglass[82] found cycles of 5–6, 10–13, 21–24, 32–35 and 100–105 years. Tree-rings have also been examined in central Europe[83] and in South-west Africa.[84] They have, however, and, especially when small material is used, to be treated with caution.

An 11-year cycle has been discovered for Japanese earthquakes[85] (plus 5·5 and Brückner's cycle), continental climates,[86] levels of Lake Tanganyika[87] and Vänern[88] (with 3·5), Finnish coasts and lakes,[89] upper Danube and other rivers[90] (with multiples, e.g. 22, 33, 44, 55, 66, 75, 133·25 and 265), and, imperfectly, for Lake George[91] (New South Wales); for temperatures,[92] especially those of winter in Europe and the northern hemisphere; for fogs in Sweden[93] and rainfall, temperature and pressure in India[94]; for growth rings in a preglacial spruce[95] and in modern trees in moist climates,[96] e.g. in California and West Africa; and for fluctuations in animal numbers[97] (see p. 137).

C. E. P. Brooks[98] states that, generally speaking, the 11-year cycle characterises equatorial regions, Brückner's cycle higher latitudes, the 11-year cycle diminishing in amplitude and regularity polewards, the Brückner cycle equatorwards. The amplitude of the 11-year temperature curve is mostly only about 0·1°C.[99]

Harvest statistics and food prices since 1678 gave a 15·3-year cycle[100] (subdivided into 5·1, 4·37 and 2·74) and precipitation in India, North America and Europe one of 15·75.[101] A 16-year period has been discovered in the temperature of Vienna and other European stations[102] and in the precipitation at Rome[103]; 18 (with 2, 3, 4, 5·7 and 11·3) in the sea-level at Brest[104]; 19 in Australian rainfall[105] and South African meteorology[106]; 21 (= 23 and 17 in succession) in tree growth in Arizona[107]; 22–23 or double sunspot cycle (see below) in Swedish and other temperatures,[108] in the rainfall of parts of the U.S.A., in the levels of the Great Lakes[109] and the Nile, in tree-rings and in catches of cod and mackerel in the Atlantic[110]; and 24 in various climatic factors in both the northern and the southern hemispheres.[111]

These short meteorological and solar periods, which in polar latitudes have a greater amplitude than the primary 11-year cycle,[112] are possibly subharmonics of longer solar periods, as is proved for example by an analysis of Wolfer's sunspot data[113]; they include many of various lengths,[114] as has been shown for solar radiation, Nile floods and tree-rings.

Among variations longer than Brückner's we may mention a 40-year cycle for rainfall[115]; recurring severe winters in west Europe and temperature oscillations in central Europe, 44·5[116]; British rainfall, 50[117]; levels of the Sea of Aral, 55[118]; winter temperatures, 69·5 and 88[119]; yearly temperatures of Rome, 76[120]; Europe's cold winters, 89[121]; precipitation in parts of North America, 90[122]; level of Lake Ngami in South Africa and of the Baltic Sea in north Germany and climatic factors in central Europe, 100[123]; trees in Arizona, 100, 150 and 275[124]; Bathurst rainfall, 108[125]; north European temperatures 110 and 220[126]; severe winters, 130[127]; New Zealand rainfall, Nile floods, tree-rings in Arizona and California and lateglacial varves in Canada, 152[128]; levels of Scandinavian lakes, 160[129]; historical droughts in the Old World, 171[130]; the postglacial dunes of the Pomeranian coast[131]; dates of the break up of the ice on the Neva at Leningrad and the Dwina at Riga, 212[132]; Nile oscillations, 213 or 225[133] (= 6 or 7 times Brückner's period with shorter periods of 5, 11, 23 and 48); certain earthquakes and climatic oscillations, 250[134]; recurring severe winters, 265[135]; Nile floods, Chinese earthquakes and sunspots, 260–280[136]; various astronomical and terrestrial phenomena, 300[137]; Russian droughts and severe winters, a few centuries[138]; Californian Sequoia rings, 300–400 and 1000[139] (with 100–150, 72, 35, 11 and 2·5); and 744 for barometric pressures, with submultiples of 372 and 186.[140]

This bewildering complexity of alleged periodicities, most of which are impersistent, induces a feeling of scepticism in their value or existence[141] which is but increased with the impossibility of finding any reasonable physical cause for them. The discovery of hidden periodicities is an extremely difficult one. Yet many mature students accept their reality and Shaw,[142] from his analysis of periodicities, concludes that many of the cycles may be harmonics of a primary cycle of 93 years and others multiples of $12\frac{2}{3}$ months or of one year or combination of the two. The no-man's-land between recent records and those of early postglacial time puts a formidable obstacle in the path of those who strive to discover the longer periods.

Relation to sunspots. Sunspot numbers, as is well known, are variable: whether this be owing to extra-solar causes,[143] such as planetary conjunctions,[144] as those of Jupiter and Venus, or to the impact of meteors[145]

("sunspot swarm") or to some intrinsic solar period connected possibly with an electro-magnetic oscillation, is not germane to our discussion. Reliable numbers are only available from 1749—they have been published for the years 1749–1900,[146] 1901–18[147] and up to 1928.[148] They show variations between 45 (1816) and 154·4 (1778) at maximum and 0 (1810) and 11 (1766) at minimum—between 1645 and 1713 there was a similar dearth of sunspots following maxima in 1625 and 1639 and minima in 1619 and 1634. Since 1900 there has been maximum spottedness at the following times[149]: 1906·4, 1917·6 and 1928·4, 1937·4 and 1947·7 and minimum spottedness at 1901·7, 1913·6, 1923·6, 1933·8, 1944·3, 1954 (?).

The length of the cycle, which is apparently best given by the minima and has been generally based upon a revised form of A. Schuster's periodogram analysis,[150] has frequently been discussed.[151] R. Wolf's table[152] shows that the cycles are not constant; since 1700 they have spaced themselves as follows: 1700–49, 11; 1750–90, 9·3; 1800–30, 15; 1830–7, 7; 1838 onwards 11·4. Schuster[153] obtained only two periods, a first with cycles of 9·25 and 13·75 years, acting successively, and a second with an 11·1-year cycle. D. Alter[154] obtained 11·37, B. Hannisch[155] 11·55 and K. Stumpff[156] 11·25. The figure commonly adopted for the recent period is S. Newcomb's 11·13[157] (Wolfer[158] gave it as 11·124, later corrected to 11·2, and planetary conjunctions give 11·178[159]), though the reversal of magnetic polarity shows that the "old" cycle of 11 years is only half of the true 22-year "magnetic sunspot cycle"[160] (Hale's period). Authorities differ as to the number and reality of significant periods which may enter into the solar graph. In addition to the 11- and 23-year cycles—the 11-year cycle, first discovered by Schwabe in 1843, has recurred 28 times since Galileo's invention of the telescope in 1610—others of 37, 68, 77, 83, 252 and 300 and even possibly as long as 1400 years have been assigned. Obviously, records of any interval of 100 years or more are scarcely yet available. While the sunspot periodicities of A. Schuster (1906) H. Kimura (1913) and H. H. Turner (1913) are apparently illusory,[161] the 11-year period is probably a subharmonic of much longer ones,[162] in particular of 89·36 years (cf. above).

The amplitude varies independently of the period by about 50% of its average value. Hence the sunspot curve should be represented, as Michelson has suggested, by a function of variable period, amplitude and phase.

Since all atmospheric movement and therefore temperature, pressure and wind distribution depend ultimately on energy received from the sun, mainly as solar radiation, it is natural that we should link climatic variations with varying solar radiation. The relationship between sunspot activity (including the number and area of sunspots, faculae, flocculi and prominences) and terrestrial phenomena, is indeed the subject of an immense literature which Hansen and Nansen have summarised[163] up to 1914 and C. E. P. Brooks[164] for the decade 1914–24.

Coincidence, early affirmed for temperatures[165] and pressures,[166] has been claimed for almost every conceivable variable, though sometimes with a phase displacement. It has been reiterated for physical, biological and demial phenomena,[167] including earthquakes and volcanic frequencies,[168] magnetic storms and intensities and electric disturbances,[169] cirrus and halo aspects,[170] temperature[171] (including that of the ground[172]), pressure[173] and atmospheric circulation[174]; rainfall,[175] e.g. of Berlin, Britain, Russia, Siberia, Dakar, Argentina, Australia and New Zealand; floods[176]; hail and thunderstorms in

various parts of the world[177]; lake-levels,[178] e.g. Lake Baikal, Great Lakes, Victoria Nyanza and Albert Nyanza, Lake Nyassa, Lake Tanganyika, Lake George and Lake Titicaca; lake-warps, e.g. Lake Sakski in the Crimea[179]; the intensity and number of cyclonic storms,[180] as in the tropics, and monsoon rains in India[181]; European vine-harvests[182]; famines,[183] commercial crises,[184] residential mortgage loans,[185] railway traffic returns in the United Kingdom[186] and historical uprisings[187]; tree-growth[188] in Scandinavia, the Baltic and North America; moist-cold periods of Brückner's cycle[189]; ice-conditions[190] about Newfoundland, Iceland, Spitsbergen and the Barents Sea and in the Baltic, Davis Strait and polar seas generally, including the Antarctic; the strength of the Gulf Stream[191] and West Spitsbergen Current[192]; the level of the oceans—the Atlantic appears to be lower, the Pacific higher during sunspot maxima[193]; and for some biological events,[194] including the fish life of the Caspian Sea,[195] the migration of birds,[196] the numerical variations of birds[197] and of certain fur-bearing animals and rodents[198], including the Canadian lynx, musk rat and snowshoe rabbits (lemmings, voles, marten and fox show a 4-year cycle, probably controlled by non-climatic factors[199]), and of crop-destroying insects[200] (though many creatures, e.g. grouse and Canadian rabbits, vary in cycles which are not primarily due to variations in solar radiation[201]). Even the integral submultiples or short period variations have their counterparts in the short periodicities of solar radiation.[202] Hale's double period has been observed in the rainfall of west Canada,[203] in the general circulation of the air[204] and in tree-growth.[205] The short period, to which various investigators have assigned lengths of 2·5–3·5 years, is given an average length of 2·33 years by H. W. Clough[206] who thinks it varies from 1·5 to 3·5 according to its position in the 11- and 36-year cycles and is connected with the mean latitude of sunspots which also varies with longer cycles. The sunspot cycle has been claimed for earlier geological periods, e.g. the Upper Palaeozoic[207] (and a 200- and a 4500-year periodicity) and the Zechstein and Eocene[208] (a rhythm of 21,000 years), as well as for the Pleistocene (see p. 144). Other periodicities have also been mentioned.[209]

But the relationship has often been questioned or rejected because the atmosphere cannot mirror promptly and faithfully slight variations in intensity of solar radiation reaching the earth[210]; temperature changes are too little to have any direct effect[211]; and meteorological phenomena do not vary in harmony or in the same ratio as sunspots.[212] Alternatively, they result from oscillations of the earth's axis.[213] Variations in magnetism and solar activity have been ascribed to a third, unknown cause.[214]

Nevertheless, the agreement, though by no means obvious or clear, seems to be too close to permit of a doubt as to the reality of some kind of connexion between the amount and quality of the emitted solar radiation, as indicated by sunspots. The causative relation, as yet undiscovered, is by no means simple or direct and the mechanism by which it is accomplished is much more complex than first believed. Solar disturbances cause an emission of atoms in swarms with consequent reactions, for example, in terrestrial magnetism when swarms are directed in such a way that they can reach the earth. The effects are probably greatest in the ozone layer, the conducting layers and auroral zone, and are only secondary at the earth's surface. Terrestrial effects, such as the lag induced by polar ice and ocean currents, also complicate them; for example, a cold year in North Siberia causes ice-accumulations between Spitsbergen and east Greenland about 4½ years later.[215]

The temperature of the earth as a whole varies inversely with the frequency and number of sunspots,[216] although the solar constant which probably averages 1·90–1·94 cal cm^{-2} min^{-1} [217] at present varies seemingly by 0·7% with the cycles of solar activity and is highest during the sunspot maximum[218]—the cycle is *c.* 23 years but is not well correlated with sunspot numbers,[219] so that the question is still debated whether solar-constant measurements as observed from the earth's surface are sufficiently accurate to reveal the period and magnitude of solar variations. This result, expressed in W. J. Humphrey's paradox of a "hot sun and a cool earth", was first announced early in the 17th century.[220] The curve of solar constant, therefore, is in opposite phase to the terrestrial curve of temperature. The paradox, which some meteorologists would deny—it depends on the region[221]—is ascribed to the earth's gravitational field,[222] to variations in atmospheric circulation through pressure conditions in monsoon and other regions,[223] to increased absorption by the upper air at times of increased sunspots,[224] and to higher radiation which augments the evaporation over the ocean and water-surfaces (thereby increasing the cloudiness and rainfall[225] and reducing the intensity of the general circulation[226]). It is also attributed to the formation of ozone in the upper air and the greater richness of the sun's rays at sunspot minimum in violet and ultraviolet rays[227] which raise terrestrial absorption—the latter are much more important than variations in the solar constant, and though chiefly affecting the higher atmosphere are reflected in the pressure and circulation patterns at the lower level. It has been sought too in Kullmer's law of the shifting of the storm tracks in accord with changes in sunspot numbers.[228] The sunspot cycle is translated into effect through the medium of cyclonic storms which have divergent regional consequences. Thus the tracks are shifted northwards in Europe and Canada and southwards in the Mediterranean and U.S.A. to control temperature and rainfall, winds and currents. Atmospheric circulation and all its attendant phenomena vary, it is claimed,[229] in unison with changes of solar radiation whose 14 periodicities, all approximately integral submultiples of 273 months, are reflected in terrestrial weather[230] (the observed variations of the solar constant obtained by the Astrophysical Observatory of the Smithsonian Institution are by some thought to be no bigger than the uncertainties of measurement or referred to defects in the methods used[231]). When solar radiation and activity are higher than normal,[232] the contrast between areas of high and low pressures is accentuated, the pressure belt in mid-latitudes is intensified, the polar anticyclone decreases and the pressure belts shift polewards. There is also a change in phase of circulation of pressure in high and low latitudes[233] and between north and south Europe.[234]

Opposed rhythms. Climatic variations are not always of the same sign over the whole globe: W. Herschel[235] noticed this as early as 1801. For instance, there is a barometric "see-saw" between Russia and India[236]; a rainfall "see-saw" between California and Florida or central North America and the Atlantic coast[237] and between the north-west fringe of Europe and the drier east[238]; and a North Atlantic temperature "see-saw",[239] cold summers in Europe usually coinciding with warm ones in Greenland and North America, and ice-rich years near Newfoundland with warm springs in Europe and ice-poor years in east Greenland, Iceland and Spitsbergen. The relationships are naturally connected with the changing strengths of the North Atlantic circulation and the Gulf Stream. Hale's period in the air-

circulation is in opposite senses in systems which belong respectively to the Icelandic Low and the Asiatic monsoon regions.[240] The minimum rainfall in continental areas is at sunspot maximum, in oceanic areas at sunspot minimum.[241]

H. Arctowski[242] examined the synchronous departures of temperature and pressure of the individual years from the 10-year average and drew lines of equal departure, designating the areas of positive departure "thermopleions" and "baropleions" respectively, those of negative departure "antipleions" or "thermomeions" and "baromeions". They travel irregularly over the earth, carrying with them conditions of high and low temperature and inducing changes of different sign. Regions of low pressure, such as Alaska, Siberia and Iceland, and of high pressure, such as the subtropical latitudes over the oceans, possess climatic variations of temperature and pressure of inverse sign in accord with the observed relation between solar prominences and the distribution of meteorological variations.[243] These "action centres",[244] from which temperature waves are propagated, have been more recently investigated.[245] They seem to move into higher latitudes with increased solar intensity.

The mean temperature in the northern hemisphere is less in winter at sunspot maximum, except over the North Atlantic Ocean and north-west and central Europe where it is higher at the maximum.[246] The sign is reversed too in North America as the tropics are approached,[247] while the ice-relationships in North Canada suggest a new centre towards Baffin Land. Stronger winds lower the temperature in the tropics and subtropics but raise it in mid-latitudes, as in the North Atlantic.[248]

Certain extensive regions during historic time experienced similar climatic changes. Others, widely separated, have either witnessed opposite changes, as Brückner[249] recognised for Burma and the Deccan, or have either remained neutral or fallen under the sway of one type or the other.[250]

The results accruing from the vast amount of painstaking work performed on these lines are, it must be confessed, meagre, disappointing and bewildering, and have merited increasingly weighty and extensive criticism. Clear and definite relationships are not established. Except for the well-known correlation of terrestrial magnetism with sunspot numbers, there is hardly a single phenomenon which is universally accepted as so related, for the figures when submitted to exact mathematical analysis give results which are usually negative. The cycles are limited and vary in their lengths and amplitudes: none can be used for prognostication, though formulas for predicting smoothed annual sunspot numbers have been developed by several authors on the basis of harmonic analysis or by numerous empirical relationships between heights of maxima, rate of rise and other factors. Yet it may be that through the control of air-pressures and atmospheric circulation[251] the meteorological elements vary with sunspots or with solar radiation and the displacement, perhaps, of the climatic belts through the latitudes. The variations do not everywhere have the same sign, the opposing variations being distributed in accord with action centres[252] as demonstrated for high latitudes.

2. *Glacier Oscillations*

Since glaciers are peculiarly sensitive to climatic variations (see p. 9), they may also be expected to have a periodicity. This is important both for its

bearing upon the general subject of periodicities (see pp. 1159, 1499) and for the light it may throw upon the cause of the Pleistocene glaciation and the fall of temperature this would require (see p. 644).

Nature of proof. Proofs of glacier retreat or advance may be gathered from ancient records, e.g. legal documents, or from natural evidence: Mercanton's labours on the Rhône Glacier[253] and O. Lütschg's work on the Mattmark Glacier[254] illustrate how the two methods can be combined.

Historical data include such things as the opening of valleys to human intercourse, and the débâcles associated with the draining of glacier-lakes. Records of this kind are plentiful, for example, in the western Alps[255] and are not quite lacking in the eastern Alps,[256] though for the period before 1600 nothing is available and glaciers are scarcely mentioned.[257] The less important events of a retreat were less liable to be recorded than those of an advance which, either by direct overriding by ice or the action of its streams, wrought havoc to habitations and lands.

Natural evidence is abundant[258]; for example *Bergschründe* and *Randklüfte* widen and become shallower and new ones are created; the bergschrund becomes a randkluft; small valley and cirque glaciers stagnate, lose their tongues, form horseshoe-shaped glaciers, break up into tandem glacierets, give way to névés or vanish completely; névés and névé patches disappear; passes become open, as in the Alps during the whole of the Bronze Age, early Hallstatt Period, 3rd and 4th centuries B.C. and later Middle Ages[259]; rock "windows" or nunataks emerge and grow, especially on steps in the valley and above the bergschrund; glaciers collapse, thin and become less convex or even concave; lateral moraines slide down and inner moraines appear; ice-flow lessens; the ends of tongues separate off, sometimes on the emergence of rock-barriers, and become dead[260]—the end of the Z'mutt Glacier, 800 m long, 120 m broad and 10 m thick, has recently separated off from the main mass; ice-caves disappear; the ice-surface becomes "tamer" or in some cases more crevassed; open crevasses heal up; confluent glaciers are dismembered—the Guslarferner has separated from the Vernagtferner, the Kesselwand from the Hintereisferner, the Schalferner from the Marzellferner, the Hofmannkees from the Pasterze—or shrink away from their lateral and terminal moraines to leave behind them moraine-covered remnants of ice or ice-free strips of land which provide interesting legal questions of ownership,[261] e.g. state, local community or private ownership; and small glacier-lakes come into existence.[262] The ice is margined by light-coloured and unweathered rocks[263] and bare-rock strips, destitute of plants, occur,[264] as on the east Alpine moraines of the year 1850, especially on the inner side where unstable slopes, occasioned by slightly moving live ice or melting dead ice, hinder growth. Comparisons of test photographs from sites accurately known, plane-table maps and lengthening distances from marked boulders or artificial cairns are corroborative.[265] Bosses of ice left on the sea-floor or partially drowned moraines forming lagoons may prove the retreat of tidal glaciers.

The time which has elapsed since the withdrawal is less easy to ascertain. It may, however, be read directly from records or gauged from the depth of channels cut through the moraines by emergent streams. It may also be calculated from the annular rings of trees growing upon such accumulations[266] and about the "forest trimline", e.g. in Alaska, British Columbia and Oregon, from liverworts whose outward growth proceeds regularly and at a rate which can be graphed,[267] or from the plant succession[268]—endolithic

lichens in the Alps are not established until after a period of about 25 years.

Proofs of positive pulsations are equally numerous.[269] Ice closes passes and cols, covers paths and camping grounds, buries small nunataks, develops dust at different horizons, and becomes steeper and more convex at its front and more crevassed and broken at ice-falls. Adjacent névés unite and tiny glaciers are born; morainic matter at the snout is diminished or eliminated or thrust up and displaced; the soil in front is cracked and beds are folded, dislocated or overturned (see p. 218); and moraines, till and fluvioglacial gravels are grooved and fluted[270]—some of these ridges, especially those which extended downstream from boulders, had probably been pressed up into basal crevasses or low subglacial channels. The ice marches over cultural formations, such as farms, houses and trees (this early directed attention to the Grindelwald glaciers), disturbs the vegetation by actual overriding, by ice-avalanches or by river-encroachment, and carries humus into subglacial streams. Dead ice is overridden or pushed up. Continuous flow may give place to block movement (see p. 118). Calving in tidal glaciers becomes more frequent and intense.

Seasonal oscillations. Seasonal oscillations in length and thickness were early observed in the Alps[271] and have been studied more recently, as in Alaska[272] and on the Rhône Glacier[273]—the latter's 20-year average gave a maximum on 25 May and a minimum on 15 October, though the length varied by only 10 m. As early glacialists observed[274] and Himalayan glaciers display most strikingly,[275] the snout swells during the winter and thins during the summer[276] in harmony with variations in temperature or precipitation. These compel an advance or recession in stationary glaciers but merely retard or accelerate the dominant variations in those which are in a state of general growth or decline.[277]

Periodic oscillations. Periodic oscillations, besides seasonal ones, were early noticed in the Alps[278] and subsequently reviewed.[279] Their study, inspired for many years by F. A. Forel, received a new stimulus in 1869 when, alarmed at the possible impairment of important water-supplies that might result from continued recession, the Glacier Commission of the Swiss Alpine Club was founded, and began work in 1874 on the Rhône Glacier, the most thoroughly investigated glacier in the world. Its reports appeared in the *Écho des Alpes* (1881, 1882), the *Jahrbuch des Schweizer Alpenclub* (1883–1890), and since 1925 in *Die Alpen*. The investigations were extended when the Sixth International Geological Congress Meeting in Zürich in 1894, following a recrudescence of the Alpine glaciers that culminated during the 1890s, created an International Commission on Glaciers which began to collect details from the whole world. Its first ten reports appeared in the *Archives des Sciences physiques et naturelles* (1895–1905), and afterwards in the *Zeitschrift für Gletscherkunde* (1906–14). Later reports, after the interruption of the Great War of 1914–18, were issued (after 1927) by a new glacier commission of the International Association of Scientific Hydrology, a unit of the Union of Geodesy and Geophysics.[280] In 1939 this commission was united with the Commission of Snow of the same body which had existed since 1933. The oscillations in the Mont Blanc massif have been particularly intensively studied during the present century[281] while those of the United States have been recorded annually in the *Transactions American Geophysical*

Union from 1931 onwards and triennially in the *Transactions International Association of Scientific Hydrology.* Reports also appear in *Zeitschrift für Gletscherkunde und Glaziologie* from 1949 onwards, and for Italy in *Bolletino del Comitato glacialogico italiano.*

Yet inaccessibility and the constant discovery of new glaciers have made detailed observations outside the Alps generally meagre; Alaska[282] and in less degree the Himalayas[283] are almost the sole exceptions. Nevertheless, data are sufficient to furnish useful generalisations.

The variation of a glacier's length within a cycle is often considerable, as in the low-lying parts of the outlet glaciers of Greenland[284]; the front of the Jakobshavn Glacier receded 11 km between 1850 and 1902 and 15–20 km between 1850 and 1931[285] and that in Eternity Fjord 300 m in 38 years.[286] The variation is usually between 100 and 150 m[287]; the Selle Glacier, for instance, retired 1·2 km in 10 years, the Pilate Glacier 0·7 km.[288] The Sefström Glacier oscillated 6 km in 14 years[289] and other Spitsbergen glaciers retreated 1500 m between 1907 and 1924.[290] The Franklin Glacier of North-East Land has receded *c.* 3 km since 1901,[291] the Stor Glacier of Sweden 138 m between 1908 and 1939,[292] and the Hornbre 3–4 km between 1918 and 1938.[293] The glaciers in Jotunheim reveal retreats of up to 500 m since the beginning of this century.[294] The Tamyngen Glacier in Turkestan Alai withdrew 2 km between 1911 and 1934,[295] the Leksir and Zanner glaciers of the Caucasus 1250 and 900–1000 m respectively between 1890 and 1933,[296] and the Murchison Glacier, New Zealand, *c.* 800 m in 45 years.[297] The Nisqually Glacier on Mount Rainier, which has the longest recession record in the United States, receded 355 m between 1918 and 1938 and 1035 m between 1857 and 1940[298] and 47 m in one year (1934), and the Easton Glacier on Mount Baker withdrew 1505 m in 31 years,[299] a record for continental U.S.A.

Movement may be great in the Himalayas where an advance of as much as 6 km has been noted in one winter and spring and *c.* 11 km in 8 days.[300] The snouts of the Biafo, Sarpo Lago and Crevasse glaciers have thinned by 46 m, 61 m and 122 m respectively in the last hundred years.[301] The snout of the Karlingerkees receded 50–80 m between 1929 and 1945,[302] and the surface of the Goose Glacier, Spitsbergen fell 40 m in 39 years,[303] the Jostedalsbrae 18 m between 1900 and 1940[304] and the Hornkees by a half between 1921 and 1937.[305] The recession is probably most pronounced in Alaska[306] (fig. 36) where one of *c.* 900 m took place in 4 years and the withdrawal in Glacier Bay was 24–26 km in 33 years, exceptionally *c.* 20 km in one year, and amounted to *c.* 97 km since 1794. Retreats of up to 20 km have been recorded since 1890 and of *c.* 11 km between 1903 and 1907. During the last two-thirds of a century the glacierised area about Muir Inlet has been reduced roughly by 35% or some 175 sq. miles (450 sq. km). By contrast, Alpine glaciers have retired up to 2 km since the middle of the 19th century; the Upper Grindelwald Glacier has retreated more than 1 km since 1822, the Rhône Glacier more than 1·6 km since 1818, the Hintereisferner 1·2 km between 1856 and 1918, the Pasterze 900 m and the Vernagtferner 3·5 km since the same maximum in the 19th century.[307] The group of glaciers of Gran Paradiso have lost on an average 20% of their surface between 1850 and 1931.[308] Certain Oetztal glaciers retreated *c.* 650 m between 1913 and 1945[309] and the Swiss glaciers have lost 25% of their area in the last 30 years. The Rhône Glacier thinned by 14·1 m between 1887 and 1907, the Vernagtferner

(near the end) by 15 m in 1888–9 and the Pasterze at the end by 16 m in 1943.[310] The surface of the Aletsch Glacier sank 52 m between 1851 and 1947.[311] The Pasterze between 1856 and 1944 lost 28% of its area and in volume 127 million cu. m.[312] The tongue of the Waxeggkees in Zillertal has completely disappeared.[313]

Plateau glaciers with their extensive firn regions react more strongly to climatic change than do valley glaciers. Since 1844, a plateau glacier in north-west Iceland, 410 sq. km in extent, has vanished and another, 600 sq. km in area, has shrunk by half.[314] The Glámujökull has practically disappeared and Hofsjökull in Lón has divided into three separate masses since Thoroddsen's time—the discrepancy between his figures for Iceland and present determinations (see p. 93) is partly due to recession. The

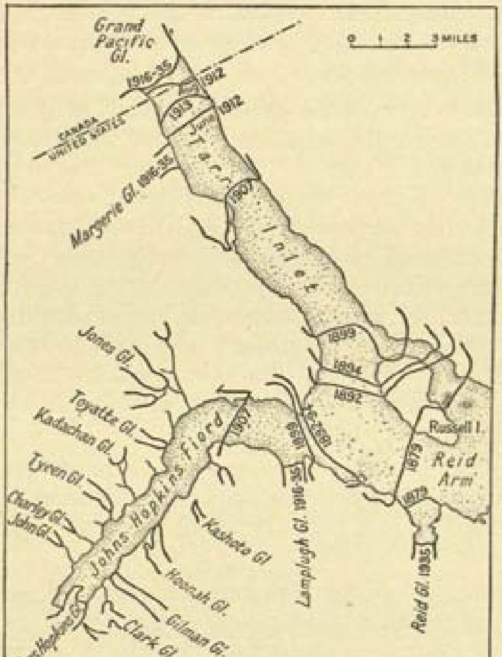

FIG. 36.—Recession from Reid Arm, Alaska, since 1879. W. S. Cooper, *G. R.* 27, 1937, p. 52, fig. 17.

Alpine glaciers have lost since the 1870s 36% of their area[315]—the glaciers of the Zillertal, Stubai and Oetztal Alps have recently lost 210 million cu. m/annum[316]—and since the beginning of the present century some of the chief glaciers of Montana have been reduced 40–75% in area and more in volume: several smaller glaciers have disappeared.[317] The glaciers on the dry south-east slopes of the Caucasus lost up to two-thirds of their surface area between 1890 and 1933.[318]

The glacier-area in the Stubai Alps contracted 28% in 50 years[319] and that in the German Alps since the 1870s by 36%.[320] The Pasterze[321] shrank between 1888 and 1927 from 32 to 24·5 km and lost between 1856 and 1936 750 million cu. m, the Rhône Glacier between 1856 and 1912 315 million

cu. m, the Suldenferner (1856–96) 85 million cu. m, the Gepatschferner (1856–1922) 143 million cu. m, the Hintereisferner (1850–1902) 200 million cu. m, and the Vernagtferner (1856–1902) 248 million cu. m.[322] The Jamtalferner lost between 1864 and 1921 42·75 million cu. m.[323] During the last 30 years, the glaciers of the Alps are estimated to have lessened in length by 10% and in thickness by 25–30 m, representing a diminution in volume of 75 million cu. m.[324] The Styggedalsbrae of Norway lost 44 million cu. m of water or one-third of its total volume between the second half of the 19th century and 1919,[325] the Stor Glacier 70 million cu. m between 1922 and 1946,[326] the Hoffellsjökull, Iceland, between 1930 and 1937 at least some 300 million cu. m of water below the level of 400 m,[327] the Barry Glacier of Alaska 1 cu. mile (*c.* 4·17 cu. km) between 1899 and 1910,[328] and the Helm Glacier, British Columbia, 5000 million cu. ft (*c.* 140 million cu. m) between 1928 and 1947.[329] Altogether, the glaciers of the temperate, subarctic and subantarctic zones have lost *c.* 10% of their total volume since the middle of the last century[330] (see below).

Length of cycle. The precise cycle is difficult to determine. Thus it is by no means easy to fix the actual year of the maximum advance or the beginning of the retreat. This is especially the case where the snout is thickly mantled with debris or snow. Additional errors creep in through short oscillations (*oscillations passagères*[331]), whether seasonal, accidental or aperiodic; by the contrasted histories of adjacent glaciers (see p. 152); and by deviations from the general trend, especially during the transition between phases. For instance, the region affected by the expanding Alpine glaciers at the end of the 19th century progressed from west to east[332]: the advance began in Switzerland in 1875 and was completed by 1898, while in south Tyrol it commenced in 1884 and in the Austrian Alps in 1890 and ceased in both areas in the early part of this century.

Secular movements probably predominate in large, longitudinal valley glaciers like the Hispar and Baltoro which are well exposed, slope evenly and gently, and are joined not too near the snout by small tributaries of similar character.[333]

Alpine inhabitants, before the 18th century, regarded the glacier-cycle as composed of a 7-year advance and a 7-year retreat.[334] This was doubted by Altmann[335] and de Saussure[336] and disproved by J. J. Scheuchzer[337] and much later by Agassiz.[338] H. Fritz[339] recognised an 11- and a 55-year cycle, and K. Faegri[340] periods of 2, 4–5, 14 and 30 years for Norway.

The sunspot cycle (and its multiples) has been demanded for glacier-oscillations[341] and Brückner's cycle has been claimed[342] for Norway (alternatively two cycles half that length), Canadian Rocky Mountains, Karakoram Mountains, Mont Blanc massif, and tentatively for Pleistocene moraines of Co. Kerry, Ireland. The Pasterze gave a period of 16·5 years.[343] Historical records[344] of the outbursts of glacier-lakes in the Alps reveal a cycle of between 20 and 45 years, with an average of the Brückner's cycle; they agree roughly with Forel's "tertio-secular period", ranging between 15 and 50 years[345]—the maxima may be 1820, 1855, 1890 and 1925. A cycle of 102–5 years (= thrice 35) has been suggested for the French Alps.[346] The Upper Shyok Glacier gave a periodicity of 44–77 years.[347]

But the behaviour of glaciers is much more complicated than this.[348] Graphs of 26 of the chief Swiss glaciers during the 19th century[349] show that

the cycle's intensity is very unequal in its numbers of periods; the Aar Glacier had one, the Rhône Glacier two and the Trient Glacier three advances.[350] It is, therefore, not surprising that Brückner's cycle has either escaped detection, as in the boreal regions,[351] where 4- and 14-year oscillations have been cited for certain glaciers,[352] or is dismissed, as for Norway.[353]

How difficult the problem is may be seen in the very different periods assigned to the very familar Alpine glaciers. Richter[354] placed the maxima in 1630–40, 1680, 1715, 1740, 1770, 1820, 1840–50, and found the 1820 maximum was the most intense and regular during the 19th century, though the 1840–50 period was general and many glaciers were largest between 1845 and 1850. Heim selected 1760–86, 1811–22, 1840–55 for the advances and 1822–40 and 1855–80 for the retreats. Forel[355] chose 1818–20 and 1850 as maxima and fixed the beginning of the decrease at 1856. Others reckon 1815–65 as one uninterrupted period with a maximum in 1818 or 1820.

Kinzl[356] has described the "early recent" or "subrecent" moraines in the eastern Alps, Swiss Alps and the Mont Blanc group of (*a*) 1850s, (*b*) 1820s and (*c*) the "Fernau moraines" of the early half of the 17th century (Grindelwald glaciers, 1600; Hintereisferner, 1680). The "1820" moraines are low walls, 1–2 m high, formed, as Richter[357] had already surmised, by a short and energetic dilatation. In contrast to the western Alps, this was less than the "1850 advance" which effaced the earlier moraines of many of the small glaciers and even some of the bigger ones, e.g. on the Gepatschferner and Pasterze. The "1850 moraines" are the most striking and often, as in such large glaciers as those just mentioned, are 1 km from the present snouts, a distance which narrows to 300 m in the smaller bodies. The "Fernau moraines" are almost as big as those of the 1850s and are noticeably rich in blocks because the ice then advanced over terrain which had long been ice-free and attacked by frost (in accord with the finding of C. Easton[358] that the winters in Europe were more open). Hence the interstadial or interglacial period of the postglacial optimum (see p. 1482) closed in the Alps with an advance which culminated in one or other of these oscillations (see p. 1496). Before 1600 historic records establish a low stand of the glaciers[359] and low air temperatures.[360]

The second decade of this century witnessed a general enlargement among an increasing number of Alpine glaciers[361] though certain glaciers, including the Aletsch and Pasterze, did not take part in it. In the eastern Alps, it was between 1913 and 1917 and in the Hohe Tatra after 1915. In the west, the advance began earlier; in Switzerland, the smallest glaciers began to advance in 1909 and by 1919 70% of them were expanding. In the Mont Blanc massif, Dauphiné Alps and Savoy all were advancing, though they were all in retreat before 1924—the Alpine glaciers[362] are now retreating annually by 2–15 m, though amounts of 15–30 m are frequent and withdrawals of 120 m are recorded. The Gepatschferner is losing annually 2·2 million and the Lower Aar Glacier up to 14 million cu. m. In the Caucasus,[363] the greatest activity was in 1850–60, 1877–87, 1907–14 and 1927–33.

The polar cycle is more obscure[364]: there are virtually no old records; modern observations are neither continuous nor so regular as in more accessible latitudes; and many glaciers are tidal and give little or no hint of their régime. There was apparently an advance from about A.D. 1550 to 1650, a recession between 1650 and 1680, a rapid advance to 1750 or 1760, a retreat until about 1790, and a maximum in the middle of the 19th century.

C. Rabot's historical retrospect[365] concerning Greenland, Iceland, Spitsbergen, Jan Mayen, Franz Josef Land and Scandinavia showed that their glaciers, after a minimum lasting several centuries, advanced greatly to a maximum during the 18th and earlier part of the 19th century and varied indefinitely and indecisively and nowhere with the importance they did in the Alps. The maximum in Iceland was 1750–60 and 1847–70—one or other of these was the maximum in postglacial or historic time[366]—, in Norway 1720 (it persisted into the 19th century) and in Spitsbergen during the first half of the 19th century. The arctic advance during the 18th century was bigger and that of the first half of the 19th century was somewhat smaller than in the Alps.

C. Rabot, J. Rekstad, P. A. Øyen and K. Faegri among others, examined the Scandinavian oscillations.[367] Rabot found an advance during the first half of the 18th century, followed by small oscillations until about 1812. A decided retreat then ensued which, with a short advance about 1865–70, lasted into the present century when another advance set in. Rekstad,[368] agreeing generally with Rabot, placed the minimum about 1700 and the maximum between 1740 and 1750. The semi-permanent snowbeds of Scotland (see p. 15) have since the 18th century responded in broad agreement with the variations of the Scandinavian and Icelandic glaciers.[369]

A cycle extending over many centuries may exist in Alaska.[370] At two former periods, Glacier Bay region was free of ice and covered by mature forests (growth rings prove some of the trees to be 250 years old) which later advances overwhelmed. These ice-free periods may have corresponded to the times when forests of giant Douglas fir (*Pseudotsuga taxifolia*), 500–600 years old, grew in the Cascade Range of Washington where now only Alpine species occur.[371]

The Himalayan glaciers seem to have had a great extension in the 19th century corresponding to that of the Alps[372]: the present tongues on Nanga Parbat are up to 600 m from these moraines. Yet the protective action of the thick morainic cover which makes the glaciers less sensitive tends to eliminate the shorter oscillations.[373]

The horizontal variations of the glaciers in recent centuries were probably more or less concurrent all over the world[374]; the *hochstands* were early in the 17th, about the middle of the 18th, early in the 19th and about the middle of the 19th century. The glaciers have been retreating from positions approximately equal to the maximum during historic time (see p. 1496), the recession being in several stages of ever-increasing intensity and interrupted by intervals of stagnation or advance.

The vicissitudes of the glaciers since the Climatic Optimum (see p. 1482) are recorded, rather obscurely, in the multiple "modern" moraines. These oscillations of historic times form part of an essentially new chapter of glacier-history ("little ice-age") which is separated by a long interval from the stadial oscillations of the declining Pleistocene glaciers (see p. 1159).

Present regression. The glaciers of the world are to-day in retreat,[375] though not always with the same intensity or without sporadic spurts of short duration and small amplitude like those which characterised the glaciers of Norway[376] and the Alps (see above) in the earlier years of this century or individual glaciers in various parts of the world,[377] e.g. Iceland, Greenland, Alaska, Himalayas and the Andes. This almost universal wasting—Baffin

Land and Ellesmere Land[378] and the interior of the ice-sheets[379] are exceptions, though this is denied[380]—which is generally impressive and in some places almost catastrophic in nature (see p. 142) is in progress, for example, in Asia[381] (Anatolia, Persia, Caucasus, Tienshan, Pamirs, Karakoram Mountains, Himalayas and north-east Siberia), the tropics[382] (Popocatépetl, Kilimanjaro, Mount Kenya, Ruwenzori and Sierra Nevada de Santa Marta)—the 15 separate glaciers of 1899 on Mount Kenya have been reduced to 10[383]—the Arctic[384] (Alaska, north Labrador, Grinnell Land, Greenland, Jan Mayen, Iceland, Franz Josef Land, Spitsbergen—in North-East Land glacier-caps are losing small ice-fields), Norway (the firn-limit has risen above many of the smaller glaciers and the recession threatens the disappearance of the glaciers), islands north of Siberia,[385] including Severnaya Zemlya, North America[386] (see below), South America[387] (Andes, Patagonia), New Zealand.[388] In the Antarctic[389] (O. Nordenskiöld[390] thought the ice in Graham Land was advancing) this is seen on Bouvet[391] and in the great outlet glaciers of South Victoria Land and is proved by the recession of Ross Barrier (see p. 179), the dead glaciers (see p. 418) and the conversion of capes and promontories into islands at the opening of the 20th century.[392] The interiors of the Greenland and Antarctic ice-sheets, where loss is by radiation rather than by convection, may not have shared in this thinning (see above).

The Alpine glaciers have been shrinking since *c.* 1850 though at different rates and by different amounts.[393] The biggest glaciers and those which face south and west or have relatively large and flat firnfields and steep tongues have withered most. The Alpine snowline, it is said, may have risen 90–95 m between 1920 and 1950 and 100–200 m or even (most improbably) 400–500 m within a century[394] (cf. p. 151). In the Sonnblick group of the Hohe Tauern,[395] ancient mining adits and workings have been newly exposed. Italian glaciers are also receding[396] (see p. 142) and many Pyrenean glaciers have disappeared.[397]

The frequency of cold winters in Europe has remarkably decreased.[398] Records in Sweden (especially in winter) reveal a rather marked rise of temperature since the beginning of the 19th century[399]: the mean temperature of winter has increased by *c.* 2°C since about 1760 and the annual temperature range has decreased. This is true for most of Europe, including Fennoscandia,[400] Denmark,[401] the British Isles,[402] central Europe and the Netherlands,[403] Russia[404] (the growing season has lengthened in Leningrad), Germany[405] and the summits of the Alps.[406] The increase is greatest in the north and interior of north-west Europe, where the boundaries between climatic regions have shifted, and lessens to the south-west where the mean annual value shows a decrease[407]—the Mediterranean region has not shared in the warming (P. Estienne, 1952). The Baltic has become milder,[408] and the times of its freezing have been altered and its salinity (because of stronger winds) has increased so that certain fish and other marine creatures, e.g. *Aurelia*, have expanded their range. Tree-rings in Scandinavia prove the present warming and climatic fluctuations as far back as the 16th century.[409] In Norway,[410] where not only the temperature but the moisture, precipitation and winds are affected, wheat is replacing oats on the lower ground, corn is ripening much higher up the hills, arable land is being extended even up to the timberline, and the treeline is rising. Lepidoptera, birds and mammals are expanding their limits in Finland, east Carelia and the

Kola Peninsula (see p. 1498) and in Spitsbergen.[411] In Russia the European hare is advancing northwards and the reindeer is retreating.[412] In Finland,[413] where the winters have diminished in length and in severity, springs have become warmer and longer and the sunshine hours have increased in number, the growth rings of trees have widened, the foliation of trees, the blossoming of plants and the ripening of fruits have become earlier, the limits of wheat and rye and of trees have moved northwards, crop values and agricultural production have increased, the equilibrium of species has been disturbed, swamp surfaces have become drier and xerophilous species have gained an increasing foothold. In north Sweden[414] bare areas are being invaded by birch and willow and tree-rings have become broader.[415] In Iceland[416] the growth of trees and vegetables is more successful and the *rústs* (hummocks) are disappearing from the *flás* (morasses) in various parts of the country. A new recurrence layer is forming in the peat-bogs of Sweden, Germany and Switzerland.[417] Denmark has been enriched by 25 species of birds, Texas by 8 and Greenland by 5.[418] The amelioration has affected the autumn change of coat of the arctic fox in Greenland[419] where the January temperatures in Jakobshavn are now *c.* 3·4°C higher than in the period 1891–1920, and where sheep and cattle are being reared in great numbers.

The temperature is rising in the Soviet Union and North Siberia[420] (see above). In North America (including Labrador), the mean temperature has risen 2°C since about the beginning of the century,[421] and the date of freezing over of the River Champlain has been postponed by 12 days[422] though the behaviour of the treeline is uncertain[423] (see p. 1498).

The atmosphere and hydrosphere of the Norwegian Sea and the Arctic Ocean have also become warmer[424] and more saline (by 0·2‰)—the temperature of Icelandic waters has risen 1°C and that of Barents Sea 1·8°C—, as have the deeper waters of both east and west Greenland.[425] This has induced a retreat of the limit of the sea-ice.[426] In the Russian sector of the Arctic the drift-ice between 1924 and 1942 was reduced by about 1 million sq. km and in 1935 an ice-breaker penetrated to 82° 41′ at the northern end of Severnaya Zemlya. In 1932 the ship *Knipowitsch* for the first time sailed round Franz Josef Land (see p. 197) and in August and September, 1940, the whole north Eurasian coast was completely free from ice.[427] There has been a reduction in the thickness of the sea-ice from 365 cm at the time of the *Fram* drift (1893–95) to 218 cm found by the *Sedov* (1937–40) and of the layer of polar water in the Arctic Sea north of Eurasia (from 200 to 100 m), and a poleward movement of the limit of the permanently frozen ground in Siberia over 40 km[428] or over an area several hundred kilometres wide along its southern fringe. The length of the coal stripping season in Spitsbergen has doubled.[429] The Spitsbergen period of navigation has increased from 95 days (1909–12) to 175 days (1930–8) and 203 days (1939)[430] and the pack-ice season north of Iceland has become shorter by 2 months between 1861–90 and 1911–40.[431] The drift across the Arctic Ocean is accelerated—the westward drift that required 25 months in the *Fram* was covered in 16 months in the *Sedov*—and the discharge of the ice into the Greenland Sea is larger.[432] Increasing Atlantic influences in the West Greenland seas have pushed southwards the southern limit of sea-ice, have weakened the Arctic water and have caused important changes in the marine fauna.[433] The sea has become warmer in the British and French region[434] (Irish Sea, English Channel), in Norway

and north Europe[435] and in the Baltic[436] (see above) where the winter ice has been reduced. There has been a warming in the Labrador Current—the temperature has risen more than 1 °C since 1912 and the number of bergs has increased (because of greater release from ice-locked fjords and bays[437])—and off east and west Greenland, north Siberia, Spitsbergen, Jan Mayen, Franz Josef Land and Newfoundland[438] and in the Bahamas.[439] The main body of water in Hudson Bay was 0·5 °C warmer in 1948 than in 1930 and was more saline by 0·5‰.[440]

The increased flow of Atlantic water into the Arctic Ocean requires a corresponding flow of polar waters southwards, viz. by the East Greenland Current and the roots of the Labrador Current, which buffers the climatic effect along their coasts (M. J. Dunbar, 1954).

There have been important changes in the marine fauna, both plankton and benthos, of the whole of the North Atlantic region[441] from the British Isles to Greenland, Jan Mayen and the Murman coast and even in the Barents and Kara seas. Cod (*Gadus callarias*), for example, has spread throughout all this region—in west Greenland in 1916 the tonnage caught was 125 tons, in 1925 1000 tons and in 1951 20,000–25,000 tons, and the occupation of the Eskimos has been transformed from one of seal hunting to fishing and the capture of whales. Arctic forms have abandoned or shortened their stay in the more southerly area, and certain fish are appearing in Siberian waters where earlier they were absent.[442] Similar changes are taking place in the North Pacific Ocean,[443] e.g. in Sakhalin and Alaska.

This warming and diminished seasonal amplitude of temperature are affecting the tropics[444] and the southern hemisphere,[445] e.g. Chile, South Africa and Batavia, though there may have been a contrary trend in Patagonia and the Antarctic[446] where ice-conditions have worsened during this century—the edge of the shelf-ice off Neu-Schwabenland is however said to have retreated at least 160 km between 1931 and 1939[447]—and in east Asia and Australia[448] and off California to judge by the changes in the fish fauna.[449] Lake-levels have fallen in east Africa,[450] e.g. Lake Victoria, Lake Nyassa and Lake Kiogo, and the surface of the Caspian Sea has been lowered.[451] The deserts have become drier[452] and may have caused a migration of birds. The absolute rise of ocean-level, from the return of sea-water locked up in the ice—some is due to sedimentation and to the upwarping of shallow seas, such as Hudson Bay and the Gulf of Bothnia—, has been during the last few decades at a rate of about 1·3 cm per century or 1 mm per annum (cf. p. 149). This relatively small figure suggests that the Antarctic and Greenland ice-sheets, which together contain about 99% of the present-day ice, have not diminished at the same rate as the smaller glaciers.[453]

The distributions of pressure and precipitation have also shifted. The subtropical high-pressure systems have moved polewards, the pressure in the north temperate zone has become higher over land and lower over water in winter, and the precipitation has increased outside the tropics.[454] Thus the annual precipitation has increased over the Arctic and north temperate zone, in Mexico, La Plata, southern India and south-east Asia, and decreased over most of U.S.A., the northern part of South America, Africa, Malaya and Australia. Winds and cloudiness, which are so difficult to measure accurately, have probably varied too.[455]

The change, which apparently shifted its centre of gravity from higher to lower latitudes[456] and has become much more pronounced in the last decades,

is partly because of a downward trend in pressure in the northern hemisphere,[457] and of an emphasising of the action centres[458] (see p. 139) and the winter anticyclones over Eurasia and North America.[459] There has been a secular increase in the stirring of the atmosphere by cyclonic circulation[460] linked with the "zonal index" over probably the whole temperate region of the northern hemisphere, including the Mediterranean region, which tends to smooth out the contrast between the warm and cold regions. The Gulf Stream has become warmer,[461] the pressure gradient has steepened, the Icelandic Low and the zonal westerlies have shifted northwards, and large masses of warm air and water are entering the Arctic about Spitsbergen[462] where the waters are more saline. The reconciliation of this view with the universal retreat may be because of altitude simply[463] or because the glacier régime in higher latitudes is controlled by conditions in the lower strata and in lower latitudes by conditions in higher strata.[464] The retreat is mainly due to the extension of the period of ablation by conduction in spring and autumn. The change in New Zealand has also been attributed to a secular movement of the high-pressure belts to the south.[465] The ultimate cause may be a change in the solar constant or activity[466] or in the transparency of the atmosphere[467] or an increase in the CO_2 of the atmosphere as a result of the great coal consumption.[468] It may also lie in the atmosphere itself or in its interactions with the oceans owing to some process which, begun almost by "accident", automatically increases in intensity until it becomes unstable or is reversed by some other accident.[469]

In North America,[470] where the climate is in general getting warmer,[471] if the retreat of the ice continues, the time is not distant when most of the glaciers in the western ranges will have disappeared. Hundreds of small cirque-glaciers have already vanished from the Sierra Nevadas during the last 60 years. Glaciers in districts close to the Canadian boundary, because of climatic factors and the lower levels of their snouts, have receded more rapidly than glaciers farther south.[472] Only the future will be able to determine the precise significance of this world-wide recession whose importance scenically and in connexion with water supplies and hydroelectric schemes needs no emphasis. The advance of the ice-front in the northern areas recently reported may have only a temporary significance,[473] though it has been suggested that there are signs of an impending reversal of conditions[474] (cf. p. 1498). Thus in south Europe and the Mediterranean region,[475] including north Africa, and in the south-east of U.S.A.[476] there has been a lowering of temperature, while there has occurred a "drying out" of the country in east Africa[477] and in arid North America[478] (see above).

Secular oscillations. Cycles of great amplitude but of unknown duration have been postulated for glaciers,[479] as for meteorological phenomena (see p. 135). Rabot,[480] for instance, recognised three kinds: "primary variations" of one or two centuries' duration, such as the Alpine advance of 1660–1720 and 1814–55; "secondary variations" within the primary ones and of contrary sign; and the problematic *periode pluriseculaire* lasting several centuries. Some Alpine moraines may denote oscillations of a higher order than Brückner's cycle[481] and secular variations of a few centuries have been postulated[482] for Italian, Norwegian and Alaskan glaciers. Greenland, on the other hand, which has experienced a recession and thinning of its outlet glaciers and margin during the last decades, has undergone no noteworthy

change during the last 1000 years. North Greenland glaciers have been stationary as far back as the Eskimos recollect and no legends are current to the effect that they were ever bigger.[483] The past retreat of glaciers may perhaps be revealed by observations of ground temperatures in the neighbourhood of glacier-snouts.[484] Beneath the glacier the ground would be warmer (see p. 561, 1344), outside it would be colder. The rate of depression of the deep ground isothermals would lag considerably behind the retreat.

Cause of oscillations. The cause of the oscillations has occasioned almost as diverse opinions as have the periodicities themselves; the ice retreated, said J. Ruskin, before the vulgarity of tourists! Several factors are involved: that these are mainly meteorological was early recognised.[485] The oscillations are accompanied by variations in the névé's extent, the *enneigement* of mountains[486] (to use Forel's term), in avalanche frequency,[487] and in the snowline which is very susceptible to climatic change[488]—in the Savoy Alps it is now 400 m higher than in 1864,[489] in the central Alps 60–80 m higher than 60 years earlier[490] (see p. 147), in Jotunheim 200 m higher[491] (see p. 646), in the Caucasus 75 ± 15 m above that of the last century[492] (*c.* 1850) and in north-east Greenland 150–200 m higher (Åhlmann, 1941).

In general terms, as de Saussure[493] discovered, a glacier's volume and its variations are a function of two variables, the one winter alimentation, mainly in the névé, the other, summer waste, affecting the tongue.[494] This has been demonstrated, for example, for Jotunheim,[495] from Alpine curves and temperature and precipitation,[496] and for the Alpine advance of the second decade of the present century.[497]

These two factors have been differently stressed. Agassiz, Charpentier, Venetz and many later investigators have regarded temperature, cloudiness and sun's radiation as the main cause[498]—valley glaciers have been likened to thermometers laid on the bosom of the earth whose fevers are recorded by the "reading" at the end.[499] The dominance of temperature is most marked in the case of ice-sheets and ice-caps since in these the upper parts are relatively the largest. The importance of snowfall is self-evident and has found frequent emphasis.[500] Only small variations are necessary to induce considerable oscillations in a glacier's length and volume. This is especially true, for example, of the Icelandic and Norwegian plateau glaciers where, because of the gentle rise of the surface, only a slight alteration in the altitude of the firnline would transfer very considerable proportions of the plateau from the reservoir to the dissipator and vice versa.[501]

Of more local significance are variations of the wind[502] (direction and velocity), local avalanching[503] and surface-debris.[504] Other causes are seismic avalanching (see p. 155), the bursting of sub- or englacial lakes (see p. 455), and volcanic activity in areas, like Iceland, where volcanoes and glaciers are contiguous.[505] A rise in temperature may explain the marked withdrawal in the Antarctic.[506]

Glaciers do not respond promptly to meteorological change: early writers,[507] arguing from the known rate of flow, thought decades or centuries must pass before it could be expressed in a terminal oscillation. In Greenland, the ice-transport must last thousands of years[508] (cf. p. 104) and the maximum extent of the Pleistocene ice-sheets, ascertained by moraines and kindred features, may have lagged long after the maximum accumulation of ice over Fennoscandia[509] and the North American centres. The lag,

however, in valley glaciers is relatively brief[510] and of the order of 2–6 years (see below). It varies with the length, size and inclination of a glacier: if the glacier is small and steep the movement starts a little earlier, if large a little later.[511] Thus, in the Alps,[512] short glaciers began to advance after 1812 and by 1817 every glacier did so. After the middle of the century, each glacier in turn became recessive, though it was not until about 1870 that all were retreating. Small glaciers began to advance again in 1909 and 1910 and medium-sized glaciers in 1912 and 1913, though the longer ones were still retreating in 1920. Rough correspondence was also found between the size of the Alaskan glaciers and the time of their advance at the opening of this century[513] (see p. 155). The shorter and steeper glaciers of New Zealand respond differently from the flatter and slower glaciers on the Canterbury side of the range.[514]

In attempting to generalise the facts of variations, one of the most serious impediments is the contrasted behaviour of adjacent glaciers. Though the sequence changes in one and the same orographic group, as Forel pointed out in 1883, tend to be in the same sense, adjacent glaciers, even if they start from adjoining firns or flow side by side for almost their whole length, are not infrequently antithetic. Such contrasted histories were early noticed in the Alps[515] and have now been described from most glacier regions.[516] The Roseg Glacier advanced this century when the adjacent Morteratsch Glacier was retreating. Even parts of a single glacier, especially if it be big and compound, may differ in phase. Thus the middle and eastern lobes of the Krimmlerkees, Venedig group, were advancing while the western lobe was receding[517], and the middle of the Suphellebrae retreated 32 m between 1899 and 1903 while one side remained stationary and the other lengthened 42 m.[518] The Presena Glacier and Gaisbergferner of the Alps have revealed similar variations[519] and Hofsjökull and other glaciers in Iceland have recently furnished further instances.[520] Baltzer[521] sought by such incongruence to explain the small advances and retreats of neighbouring Pleistocene glaciers on the Swiss Plain.

These happenings may be due to the lag in response to the climatic variations (see below), the range in altitude of the snowfields[522] or to orographic factors,[523] such as *encaissement*, narrow passes, gradient or aspect, as in the Mont Blanc group. Other factors are the relative position of the snowline and the level of maximum snowfall,[524] penetration into zones of marked ablation,[525] surface debris, possibly from landslides (since such debris makes a glacier less sensitive[526]), variations in the snow-cover which protects the lower end from melting and induces secondary oscillations,[527] avalanche falls of unusual size—of the 94 Alpine glaciers reported upon in the sixth annual report of the Glacier Commission, the only one (Boveyre Glacier, Savoy) then advancing had experienced exceptional avalanching. An occasional glacier is advancing even during the present pronounced period of retrogression. The western Alps with their steeper relief have produced more such examples than have the eastern Alps during the recent retreat.[528] In Spitsbergen, while the present warming has caused a recession of the lower glaciers, the accompanying rise of precipitation has caused an advance of high snowfields.[529] This is true also of Iceland.[530]

A tributary may advance "mechanically" (unconnected with climate), following the retreat of a trunk glacier and the removal of its damming action.[531] Several glaciers, usually separate, may cause an advance by

becoming confluent and increasing the ice-depth, the expansion being aided by the augmented snowfall induced by cooling.[532] A retreat may be occasioned by glacial capture[533] (see p. 333) or by starvation induced by an advance of other ice.[534]

The depth of the sea in front of a tidal glacier, as in the Barry Glacier, Alaska, may be reduced by morainic deposition. This, by curtailing the subsurface melting of the ice, may allow the glacier to move forward over its own sediments.[535]

Susceptibility varies with the length and velocity. It is highest in glaciers which are not restricted at the passage from firn to tongue and have large, steep firns and short, narrow tongues[536]; short sensitive glaciers record numerous oscillations.

According to Richter,[537] each glacier has a critical period of response which depends upon the shape of the outlet from the firn: the freer the egress, the more continuous is the flow. Since the shape varies, each glacier has its own time for transmitting the variation. In Richter's opinion, snow accumulates in the firn up to a certain limit when the outlet's resistance is overcome. Flow then continues until the drain on the névé exhausts the excess. The movement, once started, may go on quickly until the extra load is discharged and a state of quiescence is again attained. This discontinuous or intermittent flow harmonises with the rapidity and shortness of the advance, compared with the retreat phase in each cycle, as early discovered[538] and afterwards confirmed.[539] Thus the firn of the Vernagtferner[540] is peculiarly basin-shaped and lies behind a step 60–80 m high. So long as the basin is not filled, it continues to collect great masses of snow for a considerable time (H. F. Reid's "reservoir lag"[541]) without appreciably influencing the tongue. When full, the press of snow overcomes the friction and compels a rapid and strong outward flow which continues until the excess is relieved. The glacier, therefore, might be able to accommodate the precipitation of two climatic maxima which would escape in a single great advance. Hence glaciers, like the Vernagtferner and Suldenferner, which are constricted below the outlet of the firn, are not so well suited to determine periodicities as are broad glaciers without such constrictions.

Experiments[542] show that a certain minimum pressure is necessary to cause flow and that this, once begun, proceeds rapidly under diminished pressure. The pressure which is not uniform varies the interstitial solution[543] and consequently the fluidity of the mass. The starting of the flow is a function of rising temperature and pressure and its continuance one of localisation of maximum stresses with relief by thickening the interstitial film. A glacier's repose may be destroyed by adding unlimited lubricant either along shear-planes or at the sole so that it moves like a ship from well-greased stocks.[544]

Hess,[545] supporting Richter's view, calculated the *Stauwinkel* and "coefficient of sensitiveness". According to him, the departure from the Alpine periods which glaciers in high latitudes exhibit is due to their flat firns and long tongues, i.e. their small coefficient of sensitiveness. Great length too makes Himalayan glaciers relatively insensitive.[546] Glaciers which pour through constricted valleys from the continental ice of Greenland are in this respect more comparable with Alpine glaciers than is the Antarctic ice which is generally not so confined and therefore moves more slowly.

Richter's view has been strengthened by observations in Spitsbergen,[547] by experiments on ice-cylinders[548] and, as Richter[549] himself was aware, by the

ice-waves which travel along glaciers during periods of advance (see below). Observations on the Oetztal glaciers[550] suggest that excess precipitation may act like a hydraulic ram and that basal friction induces rhythmical oscillations of acceleration and retardation.

Forel,[551] who was the first to formulate the laws of glacier-variations, opposed Richter's view, finding confirmation in a mathematical study by L. de Marchi.[552] He believed in continuous flow and in a glacier's immediate response to increased snowfall with a progressive thickening that advanced valleywards. Variations in the firn were periodic and virtually simultaneous in all the glaciers of a particular region although, as early noticed,[553] the time a wave took to reach the snout (Forel's *retard de la période*) varied with the glacier and depended upon its length, altitude, gradient, shape of the valley, and size of the firn-basin (E. Rambert's *voyage du glacier*[554]).

Reid,[555] adopting an intermediate course, pictured a glacier as responding not by the progression of a wave but by a change over its whole surface, though the change could not show itself by reversing the phase until it had accumulated enough to carry the surface up to and beyond the equilibrium. Gilbert[556] sought to show that an alteration in the same meteorological factor or factors might simultaneously modify glaciers differing both in amount and in algebraic sign.

Waves of intumescence. Glacier variations are less simple than thought by Forel[557] who, in essaying to fix their laws, concluded that, as exemplified by Swiss glaciers, they were due to a change in volume and not in form. Refined measurements demonstrate movements affecting form as well as volume, viz. the passage of waves of intumescence (Finsterwalder's *Schwellungswellen*[558]) through the mass.[559] The greater accumulations above the lowered firnline and the lengthened stream-lines necessarily make new additions below this line. The waves travel down the glacier, widening it to dam glacier-lakes, as in the St. Elias Range, Alaska, and ultimately lengthening it. At the snout, live ice overrides dead ice[560] and upper layers move over lower ones along definite planes. A tributary may override a trunk glacier or cascade on to it; it may also deform the median moraine.[561] The advance of the end, heralded by a thickening of the upper regions, is due to the arrival of the wave. A small expansion of the transverse profile propels the snout disproportionately and causes the glacier to "run away"—advances of the snout by amounts of up to 76 m per day have been recorded.[562] This was shown by Forel and Richter. S. Finsterwalder, formulating it mathematically, demonstrated that oscillations took place within a space which was limited by the surface of a stationary glacier at its maximum height above and minimum position below. The longitudinal profile becomes convex upwards during an advance but flattens and becomes even concave during the receding phase; the surface then sinks, the margins fall in, glacier-caves collapse and crevasses become finer and shallower.

The maximum length and greatest cross-section of the tongue are clearly not simultaneous.[563] The upper part may be already declining when the wave attains the snout as noticed by A. Escher in 1841 on the Aletsch Glacier and by later workers in the Arctic and on the Vernagtferner.[564] As a corollary it follows that a glacier does not rise to the level of its lateral moraine at one and the same time.[565] On large glaciers, several waves of short duration, such as the yearly oscillation, may be on their way to the tongue.

The enlargement of volume during the passing of a wave[566] is very con-

siderable, as in the Savoy, but becomes less towards the close of a retreat. The Glacier du Tour, for instance, increased annually between 1911 and 1920 by 8·5 million cu. m and the Allalin Glacier between 1915 and 1929 by 4·4 million cu. m.[567]

The wave, much magnified in glaciers with narrow tongues contrasted with the firn-basins, swells towards the snout[568]: in the higher parts the rise is not a wave but a movement carried by the whole mass.[569] Its velocity of propagation, as analysed mathematically,[570] is many times the rate of flow; it was 20–150 times on the Hintereisferner[571] and amounted in Alaska to as much as 24 km/annum or 60 m in one day, i.e. a movement which can be detected in one minute's observation. The Black Rapids Glacier of Alaska—the "runaway glacier"—rushed forward *c.* 5 km in 1936–37. Nevertheless, the flow is raised simultaneously.[572] Thus the advancing Vernagtferner at the beginning of this century,[573] which advanced explosively in 1848 by 500–2000 m, increased its flow to 280 m/annum though the average profile rose only 1 m—a measurement in 1845 gave a rate of 1·9 m/second.[574] A glacier advance is, therefore, as in the Oetztal,[575] heralded by a big addition to the rate of flow.

This furnishes the key to the observation, often made, that the flow is higher during an advance than during a retreat, and explains why periods of advance are usually shorter, probably about twice as short,[576] as those of retreat. Incidentally, it shows too that we can only have a true conception of the flow of a particular glacier when measurements extend over a sufficient number of years to embrace both an advance and a retreat phase. It reconciles too the apparently conflicting daily velocities which have been recorded from one and the same glacier at times widely separate, such as the *c.* 2 m and 12–21·5 m recorded for the Muir Glacier, Alaska.[577]

These waves accord with F. Höppler's experiments[578] which showed that the apparent viscosity was a function of the applied shear stress, the temperature and the crystal orientation, and that at a constant temperature of -1°C an increase in the shear stress from 10 to 63 kg sq. cm reduced the apparent viscosity by a factor of 200.

Seismogene oscillations. Alaskan glaciers, including the Malaspina Glacier and those of Yakutat Bay, provided a special case of waxing and waning at the opening of this century.[579] During their spectacular sally (spurts of up to 2 km in 19 months were registered and in some cases, e.g. the Altrevida, Haenke and Variegated glaciers, the complete cycle was accomplished in *c.* 1 year) the ice-surface was profoundly transformed by crevassing and massive breaking (the "troubled surface of a glacier flood"); the glacier ends were greatly thickened (at least 60 m on the Altrevida Glacier[580]) and expanded in the piedmont bulbs; marginal valleys were closed up; and alluvial fans were overridden. In the succeeding 10 years icebergs became far more plentiful. This phenomenal advance was followed by retreat and stagnation[581] that differed in abruptness and intensity from those of a normal cycle of precipitation.

The advance resulted probably from the severe and repeated earthquakes of September, 1899, which took place along a series of lines parallel to the shores of Yakutat Bay fjords.[582] The "seismic avalanches" shook down in the gathering grounds unusual quantities of snow. The different behaviour of the various glaciers to the disturbance in the budget balance was related to

their lengths, the local intensity of the shocks, the quantity of snow available for avalanching, and the steepness of the rock-walls.

In Alaska the lofty mountains, steep slopes and heavy snowfall were especially favourable. Seismic disturbances have, however, affected avalanche frequency in the Alps,[583] e.g. Biez and Giétroz glaciers, and in the Caucasus,[584] and have given rise to large block moraines in the Canadian Rockies[585] and to excessive debris on some glaciers in the Pamirs.[586] They may also have been responsible for the dead ice of certain areas[587]; for Alpine block glaciers[588]; and for the large oscillations[589] in North-East Land (possibly) or in regions like the Himalayas and Karakorams which are still subject to faulting and seismic disturbances,[590] or Turkestan[591] where the steep and high walls above the glaciers are especially favourable (see p. 33).

Seismic avalanches are a thoroughly justified adjunct to the normal theory of glacier oscillations. Since their process is strictly self-limiting, it is quite erroneous to extend their application, as has been done,[592] to explain the Glacial period itself or its oscillations, or to attempt by their use to reconcile the multiple glaciations of the Alps with a single epoch in north-west Europe.

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(*B*) *SEA-ICE*

Chapter VII

TRANSITION ICE

The Antarctic continent, an ice-sheeted land girdled by a wide sea, has a much severer climate than the Arctic as isotherms reveal[1]; within the 60th Parallel, its summer is more than 10°C colder. The polar position of the land-mass, its great average height, the low vapour content of the atmosphere which reduces clouds to a minimum, and the high reflecting power of the snow and ice conspire to make the Antarctic a vast refrigerator. By contrast, the expansive land-locked sea of the Arctic is penetrated by relatively warm ocean currents from the south and its glacierised lands are mostly islands and generally too small and too low to influence the climate markedly.

North of 60° N. Lat. there live more than one million human inhabitants; south of 60° S. Lat. there are none, nor is there a flowering plant or a single land animal larger than an insect. The Antarctic cold has also important glaciological consequences. It depresses the snowline, slows down the ice-flow, and retards the conversion of snow into ice—Antarctic glaciers and bergs have small granules (see p. 33) and have undergone few structural changes. It compels the ice to spread as far as or even beyond the limits of the land—in the Arctic similar conditions are only to be found in Franz Josef Land where low temperatures are combined with ample supplies of moisture from the North Atlantic "Low" (see p. 84). It has generated special kinds of ice: ice-domes on the small islands and coastal shelf-ice and floating glacier-tongues. While, therefore, the Arctic presents a strong contrast between ice and water which turns rapidly in favour of the water at the ice-edge, the Antarctic presents a transition from ice to oceanic conditions.

Shelf-ice, floating ice-tongues and ice-foot should be considered with the land-ice to which they more properly belong (see p. 72). It is, however, more convenient to treat them as neighbours of the sea-ice with which they commonly associate.

Shelf-ice. The term shelf-ice which O. Nordenskiöld[2] introduced for the ice of King Oscar II Land (Ferrar[3] called it "piedmont aground" and "piedmont afloat" and others[4] *Meergletscher*) covers the most important of the transitional masses, transitional in their position and in the fact that they float like sea-ice but like land-ice stay in place or move very slowly. Some authorities,[5] including the United States Board on Geographic Names,[6] prefer the name shelf-ice, though other glacialists[7] prefer the older name barrier which has always had an areal significance, e.g. Ross Barrier and Filchner Barrier. Except for these two the term shelf-ice is now generally adopted.

The total area of the Antarctic shelf-ice, which varies from year to year, was calculated at 930,910 sq. km in 1948[8] when the major units had the following areas (in square kilometres); off Neu-Schwabenland, 16,272; Ingrid-Christensen shelf-ice (Mackenzie shelf-ice), 12,672; West Ice (Drygalski shelf-ice), 28,944; Shackleton shelf-ice, 39,024; Lady Newnes shelf-ice,

4610; Drygalski foreland-ice, 2736; Ross Barrier, 487, 826; Sulzberger shelf-ice, 12,672; Getz shelf-ice, 20,004; Wilkins shelf-ice, 16,992; George VI shelf-ice, 28,224; Wordie shelf-ice, 432; Nordenskiöld and Larsen shelf-ice, 102,960; Filchner Barrier, 165,454; and Stancomb-Wills shelf-ice, 2888.

The Ross Barrier or Shelf-ice, described by Sir J. C. Ross[9] in 1847, was the first example to be discovered. This most distinctive of Antarctic features[10] (fig. 37) stretches in Lat. 77–78° S. from South Victoria Land to King Edward VII Land, a distance from east to west of some 800 km. It

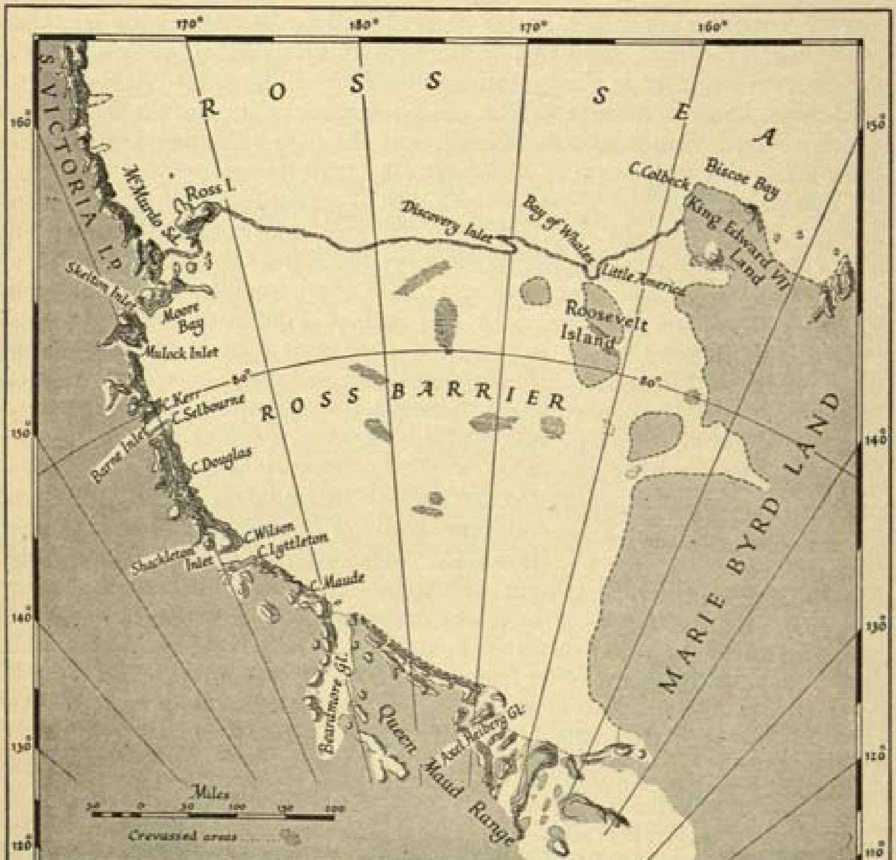

FIG. 37. Map of Ross Barrier. F. Debenham, *G. J.* 112, 1948, p. 197, fig. 1.

extends backwards to 85° S. Lat. through almost the same distance (650 km) to an inner boundary which Scott, Shackleton and Amundsen delineated in the west (South Victoria Land) and south (Queen Maud Range) and R. A. Byrd and L. M. Gould discovered in the east where King Edward VII Land, Rockefeller Plateau and Marie Byrd Land hem it in.[11] Its area is roughly the size of California or France, or of the North Sea from the Straits of Dover to a line joining the Orkney Islands to Bergen, i.e. 700,000 sq. km or over 250,000 sq. miles.[12]

Its surface, as the *Discovery* (1901–3) first proved, is a featureless plain of dead monotony; the average height has been estimated at 52 or 70 m and over the 700-km stretch south of Framheim is only 60 m.[13] On the north it breaks off in a straight and regular white cliff which forms the southern navigable limit of Ross Sea and so provided a "barrier" to Ross in his quest for the South Magnetic Pole. The Barrier is dissected at its edge by vertical joints[14] due to tidal strains and alternate expansion and contraction induced by temperature variations.

The shelf-ice, as its face reveals, consists of compressed snow and stratified, compact névé, whose grains grow and interlock downwards more intimately[15] as the result of mechanical settling and compaction without melting—there is in consequence no orientation of grains. The density increases downwards and attains a figure of 0·5–0·6 at a depth of *c.* 6 m[16]—the density in the Maudheim shelf-ice increased logarithmically from 0·45 just below the surface to 0·82 at 60 m and 0·88 at 100 m, the increasing density below 60 m being entirely due to the compression of the air bubbles. The strata are horizontal and even, though the upper ones curve slightly to conform with the surface inequalities of the drifting snow.[17] The strata vary considerably in thickness in short horizontal distances. Save near the pressure-ridges in which the firn has locally been converted into glacier-ice, no solid ice is visible above the waterline. Yet since the shelf-ice is partially fed by outlet glaciers e.g. those of South Victoria Land (Skelton, Barne, Shackleton, Beardmore, Reeves, David and Ferrar glaciers), its submerged base may be ice (see p. 171). Its relatively light structure is confirmed by the density of its accessible parts and by soundings around stranded bergs which have broken away from it.

Its northern strip is afloat[18] and rises and falls with the tide. Tidal cracks and hinge-lines extend some distance back from its face, as around Mount Terror and White Island and in the region of the Bay of Whales[19]—along the tidal cracks parallel with the edge sea-water freezes and, the process being repeated, causes "hay cocks" or domes of ice to form at the "breathing holes" from the warm air and vapour from the freezing sea-water.[20] Ridges or crevasses are absent, unless the Barrier locally rests on the sea-bottom, as between latitudes 81° and 82° S. on the 175th meridian, or where the impact of the outlet glaciers has piled up huge pressure-ridges which grade into low rolls and open folds concentric about them and 3–5 m high[21]—the ridges in front of the Beardmore Glacier extend for fully 32 km into the Barrier.[22] Moreover, its movement (see below) is large compared with the measured rate of motion of the glaciers and is not differential in depth[23]; its face has rapidly retreated since 1841 (see p. 179); the temperature rises in the fissures and shafts as the water under the ice is approached[24]; and currents flow in and out beneath the ice.[25] This conclusion is strengthened if the Barrier's height is compared with the depth of Ross Sea immediately below. The shelf-ice is extremely tenuous compared with its area (like "a sheet of paper floating upon water"). Its height varies from place to place and time to time between 2 m and 85 m and averages perhaps 24 m, giving a total thickness at the seaward edge probably much less than the figure of 220 m mentioned in the Shackleton memoir,[26] though this may be more than equalled in the south[27] —G. Taylor judged its average thickness to be *c.* 300 m. Seismic soundings have given figures of 160–200 m.[28] In general the height near the edge is 10–15 m lower than that, say, 1000 m

from the edge, the difference being due to the more rapid melting of the base of the barrier at the edge.[29] The shelf-ice may be afloat over an area some five times that of the southern part of the North Sea, though it is grounded about 10 miles (16 km) south of the edge in the area of the Bay of Whales[30] (fig. 38; pl. VB, facing p. 112).

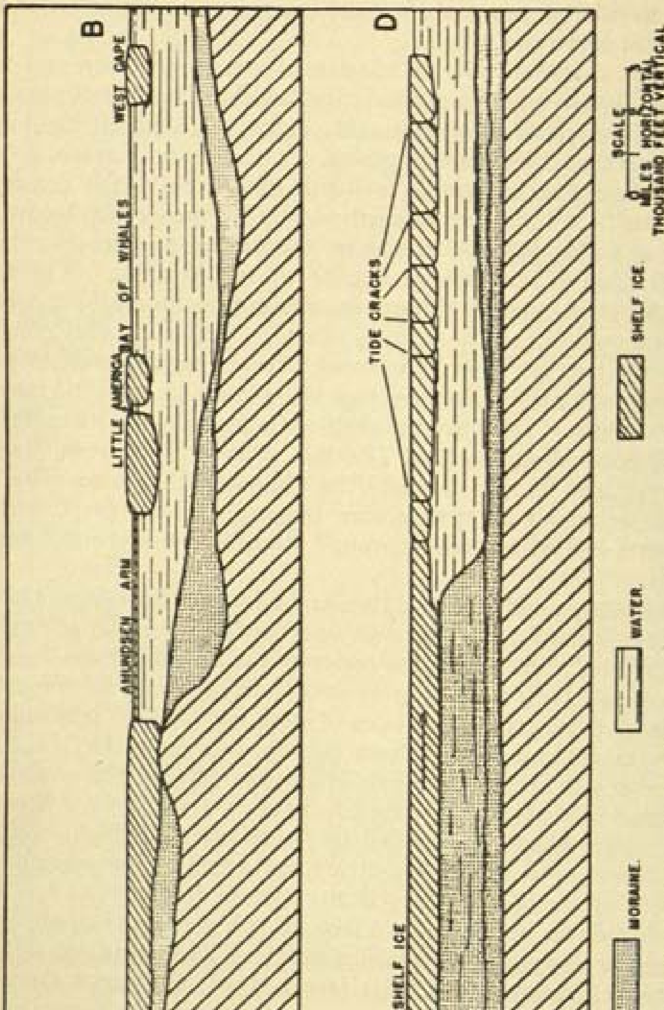

FIG. 38.—Vertical sections of Ross Barrier and morainic material along south-north lines west of the Bay of Whales. T. C. Poulter, 1909, p. 380, figs. 12 and 13.

That the shelf-ice is moving forward is proved by pressure rolls and crevasse systems of the piedmont glaciers and by the Barrier-face which has not changed very materially this century, although icebergs are breaking away continuously. Yet its rate of movement over the almost frictionless surface of the sea beneath has rarely been ascertained. Off Mina Bluff,[31] it exceeds 1 m/diem (or 356 m in $13\frac{1}{2}$ months) and may be many times this since the measurement was probably not taken in the line of maximum flow. F. Debenham[32] deduced a figure of 4–5 m/diem for much of the face by comparing Scott's survey of 1902 and 1912, while Hess,[33] from published maps, found it varied between 1·7 m and 2·9 m, though the total width of the outlet glaciers compared with the Barrier's height gave 5 m. Measurements around the Bay of Whales prove a daily flow of 2–4 m.[34] A velocity of this order, also demanded apparently by the number and size of the Antarctic bergs which calve here, may be explained by the relatively high temperature of the base,[35] estimated at *c.* -2°C where sea-water laves it.

The flow in the north-east is severely handicapped by islands and reefs, proved by seismic soundings,[36] including Roosevelt Island which rises about 30 miles (48 km) south-east of Little America to *c.* 260 m and is capped by *c.* 168 m of ice[37] and causes the Bay of Whales to form in its lee. The flow is compounded, in roughly equal proportions,[38] of the thrust of the active outlet glaciers and of the movement due to the flattening and outward extension of a thick mass of ice under its own weight. The effective water circulation below may not be without its influence.[39] The ice-aprons of the outlet glaciers are doubtless prolonged as ribs running for relatively short distances through the lower level of the shelf-ice and forming thin floating piedmonts, knit together and masked by surface snows derived from wind drift and local precipitation.[40] Yet while the shelf-ice towards its seaward edge may overlie the buried ribs or brackets, since flat-topped névé bergs, with a base of land-ice, have been sometimes seen,[41] it is more likely, in view of the ever-increasing load of snows and basal melting, that the whole of the north central section is a secondary product[42] composed of névé only. This likelihood is reinforced by the comparative absence of large erratics from the floor of Ross Sea.[43] Pits sunk into the shelf-ice show that 5·5 m of firn accumulated between 1940 and 1947, that recrystallisation has not occurred to a depth of 11 m and that differential motion to this depth does not exist.[44] The average density of 0·84 may be reached at *c.* 30 m by compaction. At a depth of 41 m the temperature was nearly constant at -22°C.[45]

Origin. The history and origin of the Ross Barrier, as of other shelf-ice, are not definitely known. A relic from the Glacial period[46] and an offspring of parents long since disappeared, it may owe its existence to snows drifted upon an original sheet of sea-ice or a piedmont of glacier-ice[47]—in barrier-bergs the stratification of snow layers or bands extends downwards to about 30 m from the surface[48]—or to floating ice-tongues expanding and coalescing in the back of an indentation to form rafts, cemented by interstitial sea-ice and intervening snows.[49] The yearly excess of snowfall over surface-loss averages 33 cm of hard snow or 20 cm of solid ice.[50] Shelf-ice may arise if snow falls during successive seasons on bergs and sea-ice, as in the case of West Ice (*Westeis*) of Kaiser Wilhelm II Land (see below), or on sea-ice alone,[51] e.g. the Weddell Barrier, either on a very cold lee shore whose hinterland is not extensive enough to provide glacier-ice (as in the lee of Posadowsky Bay in Kaiser Wilhelm II Land), or on a coast like Graham Land

(Larsen shelf-ice) which is bordered with islands and shoals which give shelter from sea-ice. Similar *points d'appui* are associated with Shackleton shelf-ice, King George V Land, and the shelf off Ellesmere Land (see below), and occur under Ross Barrier and produce pressure waves on their inner side and crevasses round their edge. This view was put forward for the Antarctic bergs by early explorers,[52] including Hooker[53] who regarded Ross Barrier as solid pack-ice which was covered with snow from year to year and gradually sank. E. Philippi[54] criticised it because transitions to sea-ice are unknown and the *Gauss* Expedition found that basal melting limited the thickness of such ice.

Drygalski's *Westeis*[55] is an agglomeration of bergs and pack-ice, 6000 sq. km in extent. Stranded on shallows and densely packed, it has been welded together by and shrouded in drifted snow. Its uneven surface, in places chaotic and stagnant, terminates in a vertical face up to 20 m high. Soundings and perceptible vertical movements show that it is afloat. It may be a Pleistocene relic which has thinned and begun to swim but has been prevented by shallows from floating away.[56]

Drygalski's *Blaueis*[57] is made of bergs and drift-ice, rounded, smoothed and polished by the wind and cemented together by freezing sea-water and drifted snow. Its outer and more humid zone is channelled by melt-waters and passes through a transition stage (*Blau-Murbeis*) into porous ice (*Murbeis*).

The shelf-ice of the Gaussfield differs from Ross Barrier in its passivity (it moves only by external force). It is not replenished by surface snows but is situated in a region of ablation.[58]

Filchner Barrier,[59] discovered and described by W. Filchner occupies the southern portion of Weddell Sea south of 69° S. Lat. Its edge is at least 450 miles (725 km) long and 10–25 m high. Deep water, averaging 300 fathoms (*c.* 550 m) and immediately beneath, proves that it floats. The Larsen Shelf-ice (Lower Ice Terrace of O. Nordenskiöld) occupies the western part of the sea east of Graham Land.[60] It extends as a belt 16–160 km wide from near the tip of Graham Land (63° 45′ S.) for about 1000 km to 75° S. Lat.: smaller ice-shelfs occur to the south. The narrowness of the shelf in this direction is probably partly due to lessened precipitation and partly to the sudden change in the direction of the east to west current in the Weddell Sea at Nantucket Inlet. The edge of the Larsen shelf-ice seems to have changed little, if at all, in the last 50 years.

Shelf-ice, besides the masses just mentioned, has been observed[61] near King Oscar II Land; by Cook and Biscoe west of Kemp and Enderby Lands; north-east of King Edward VII Land and north of Neu-Schwabenland; on Peter I Land; in King George VI Sound, Graham Land (including Wordie Shelf-ice), where it is *c.* 915 m thick; in Western Dronning Maud Land where it is *c.* 190 m thick and moves *c.* 303 m/annum (see above); and as the Stancomb Wills Promontory on Coats Land, a remnant of a vast sheet that formerly covered this region.[62] While the precise extent of all shelf-ice is not certainly known, it is next to the inland ice the most extensive of all ice-types, its distribution being related to latitude. In the Antarctic where the shelf-ice has provided stations for several expeditions, e.g. Whale Bay, Ross Barrier (1911, 1929, 1934, 1940), Shackleton Shelf-ice (1912) and Queen Maud Shelf-ice (1950–51), it may cover about 1 million sq. km[63] (see p. 167). In marked contrast, shelf-ice is almost unknown in the Arctic. The "glacial fringe" of Ellesmere Land and north Grant Land,[64] which offers certain

analogies and may be classified with it, forces its way over a shallow and fairly level sea-bottom and is parted from the polar pack by a shore lead or tidal crack. It gives birth to low floebergs, up to 30 m thick, to palaeocrystic ice, and to "ice-islands" which, many kilometres in length and up to 75 m thick, have been recorded in considerable numbers especially in recent years.[65] They have gently corrugated surfaces with scattered patches of mud, sand and boulders (derived by freezing at the base), originate by breaking off at high tide and with strong winds, and drift at the rate of about 1·1–1·2 miles (*c.* 500 m) per day over the Arctic Ocean and into the several channels of the Canadian Archipelago. Other ice-islands, up to 700 square km. in extent, have been seen recently by Russian explorers and may explain the "lands" reported by earlier explorers, such as Sannikov Land and Andreyev Land. Shelf-features occur in Novaya Zemlya[66] and others possibly in Franz Josef Land[67] and on the north coasts of Severnaya Zemlya.[68]

In Yökel Bay, north-east Greenland, between 78° and 82° N. Lat., a floating mass, 40 km broad, covers several thousand square kilometres. It rises and falls with the tide and moves forward into the sea as its crevassed hillocks and wavy surface prove.[69] Unlike the Antarctic shelf-ice, which it superficially resembles, it consists of true glacier-ice.

Floating ice-tongues. Other features of the Antarctic littoral are the "floating ice-tongues".[70] Shackleton Shelf-ice,[71] the biggest of these long, narrow and subangular peninsulas, probably floats for 160 miles (*c.* 260 km) of its length and extends 200 miles (320 km) from east to west and 180 miles (*c.* 290 km) from north to south. That the tongues are afloat is shown by their tidal rise and fall, by soundings, by their steep and high sides, by their rare crevasses, and by cracks in the adjacent sea-ice.[72] Their surface which may be at 45 m is convex, except at the distal end which is horizontal and afloat. The plan is triangular, due to lateral calving,[73] but is rectangular if the flow is considerable, as in the Nordenskiöld and Mackay Ice-tongues.

The tongues are maintained by direct snowfall and by glacier thrust from behind. The single velocity that has been measured shows that the Mackay Ice-tongue[74] moved forward during the summer 115 cm/diem: the rolls in the tongue may be due to this movement. Since the tongues are floating and rest on virtually frictionless bases a negligible slope causes ice-flow.

The chief examples in the Antarctic[75] are Termination Ice-tongue at the western end of Wilkes Land (in 1931 it had broken up into a number of grounded bergs[76]), Shackleton Shelf of Queen Mary Land, Nordenskiöld (37 km long) and Drygalski (64 km long) glacier-tongues of South Victoria Land, Sir John Murray Tongue, Dugdale and Denham Ice-tongues, Mertz and Minnis glacier-tongues of Adélie Land (80 and 140 km long respectively); and the ice situated west of Vahsel Bay off Prince Regent Luipold Land.

The floating terminations of several glaciers may unite into "confluent ice".[77] These floating equivalents of the piedmont glacier bear the same relation to this that the floating ice-tongues do to the expanded-foot glaciers. They are relatively unimportant and practically restricted to the Antarctic, arctic examples being known only from Spitsbergen.[78]

Although glacier-ice in the Antarctic descends into the open sea in but few places, every vigorous glacier (and many which are not) is able to maintain a floating extension. But these floating ice-tongues, like shelf-ice, are rare in the Arctic; for the meagre precipitation, more rapid shore-currents and

positive summer temperatures are inimical.[79] The latter, moreover, prevent the healing of the crevasses which open between the floating and fixed positions. Hence, tongues of this kind can only exist in arctic glaciers which are not crevassed in this position, i.e. those which have a low gradient and discharge into fjords, densely packed with permanent sea-ice.[80] Examples are the Ryder, Steensby and Petermann glaciers[81] (the Petermann's outer 40 km are afloat although the edge is only 2·5 m high), the floating ice-mass in the Nordenskiöld Fjord in north Greenland[82] and the hammer-shaped masses in the ice-fjords of Severnaya Zemlya.[83] The Turner Glacier,[84] Alaska, may belong to this category.

Ice-foot. The ice-foot was noticed in polar regions by such early explorers as J. C. Ross, E. K. Kane and G. Nares. It has been described and pictured from the Antarctic[85] and Greenland[86] in which country it was known to the Danish colonists as *isfod*, a term which Kane[87] first translated into "ice-foot". The type passes into O. Nordenskiöld's "ice-foot glacier",[88] Arctowski's "slope glacier" or "suspended slope glacier",[89] Gourdon's "piedmont glacier"[90] and Holtedahl's "strandflat glacier" and "Antarctic ice-mantle type".[91]

This low, flat terrace skirts the polar coasts more or less continuously just above sea-level[92] (pl. VIIB, p. 192). It surrounds islands and stranded bergs and fringes quiet glaciers which are aground and the edge of the Antarctic ice-sheet.[93] Its ribbon of ice, firmly frozen to the ground, follows the coastal undulations, serving as an excellent and favourite winter highway for the Eskimo's sledge journeys. Its level top marks the highest tide of the year[94] and rises to 3 m or more. It is especially broad on shores which are protected, as by off-lying islands, from the crushing activities of sea-ice. In Kane Basin it is 100 m or even several kilometres broad.[95] It is very narrow on steep and rocky cliffs which rise from deep water; in many localities in north Greenland it is only 0·5 m broad and on vertical and very exposed coasts may be quite wanting. The outer edge falls steeply or vertically towards the sea-ice and coincides with a tidal crack along the line of ebb tide. On gently shelving shores, where the terrace is very wide, the transition to floe-ice is scarcely perceptible and there are many tidal cracks.

The ice-foot approaches glacier-ice in structure but is distinguishable by its larger air inclusions, its salinity and often by its regular stratification.[96]

This transition type is of composite origin.[97] "Drift ice-foot" is fed by snow drifting from the land and is greatest on lee cliffs and glaciers. Being the last to grow and the first to disappear it is the least persistent. "Spray ice-foot" grows from spray and sea-water, notably in heavy weather, when islets may be capped with masses of ice soaked by breaking spray. It is very general in the Antarctic, e.g. in South Victoria Land,[98] especially windward of projecting points.

Calving floods wash blocks of sea-ice on to the ice-foot and winds and tides thrust them over one another.[99] Drifted snows and frozen spray fill and cement the intervening spaces so that the whole is consolidated and smoothed over. This "pressure ice-foot" originates in deep bays or on coasts exposed to the pack. Pressure from the latter and the action of the sea produce modifications.[100]

Tidal ebb and flow also build up an ice-foot from low-water mark, each receding tide leaving its congealed encrustation.[101] This "tidal platform

ice-foot"[102] is accreted roughly through the tidal range and is best developed where this is considerable, as in Baffin Land. It may line a quite vertical cliff and may be sensibly added to on shallow coasts by anchor ice at its lower edge. On the free and open coasts of the Antarctic its breadth, governed by coastal configuration, may be 90 m.

"Storm ice-foot",[103] which is exceptionally high and overhangs seaward, forms during heavy swells in the Antarctic but is unknown in north Greenland. "Melt ice-foot"[104] projects from bergs as a submarine extension.

Auxiliary aids in building up the ice-foot are mists and ice-crystals which freeze on to rocks. Breccia and pack-ice material are important additions on bergs.[105]

The ice-foot preceeds the formation of sea-ice and survives the disappearance of this ice but is only permanent if the summers are cold, as in north Greenland. It melts rapidly in spring[106] by waters from the land which erode deep gutters along its edge, by warm sea-water which undermines it and by the sun's rays, aided by grit blown out by the wind.

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CHAPTER VIII

DRIFT-ICE

Drift-ice may be usefully subdivided into icebergs and sea-ice. The first is land-ice which is discharged into the sea (very subordinately it embraces river-ice); the second arises when sea-water freezes.

1. *Icebergs*

Arctic bergs are usually born in glacier-bearing fjords; H. Rink[1] distinguished such ice-fjords (*iisfiorde*) from stream-fjords (*strömfiorde*). Their appearance is rugged and irregular[2] since their sides are crevasses or joint planes; castles, churches, domes, spires and columns are among the words commonly used in describing them. Several subtypes may be recognised.[3]

Manner of calving. Calving depends for its rate and amount upon a glacier's velocity, thickness, slope and degree of crevassing, and upon the configuration of the sea-bed. Since these remain more or less constant, big bergs in the case of a particular glacier tend to break loose at approximately the same place: the greater ice-flow of summer and the greater freedom from sea-ice produce a maximum productivity at this season. The manner of calving is, however, imperfectly known since the event has been witnessed only occasionally and by few scientific observers.[4] According to the earliest conception, a glacier pushes out into the sea until its outer part, losing its support, breaks off by its own weight: Rink's observations on the Jakobshavn Glacier[5] showed this to be erroneous. The floating termination is more and more buoyed up, the upthrust producing tension and consequently basal crevasses. Small pieces are calved if, as on the north and west coasts of Novaya Zemlya,[6] a glacier flows into a shallow and gradually shelving fjord.

Rink's calving by upthrust is implied in the Greenlander's word *igarpok* ("inclined backwards"): many glacialists, including E. v. Drygalski, R. Hammer and A. Helland, have recognised upthrust as a factor. But opinions differ as to whether a glacier calves as soon as it leaves the bottom or only after it has moved a further distance. While Rink[7] himself favoured the second view, Drygalski[8] thought that soundings and old water-lines in the ice-face proved the alternative. In his opinion, a glacier extends in summer slightly beyond this critical position but pack-ice prevents this in winter—an advance of 2–3 km is, therefore, possible when the winter-ice is dissolved.[9] K. Steenstrup also held that a glacier's motion, aspect and crevassed and vaulted surface were irreconcilable with floating.

The crucial question in the discussion is whether newly calved bergs do or do not project above the glacier's snout: projection, frequently affirmed,[10] is as often denied.[11] Some, doubting if it has been seen or proved, deem it possible only where, as in the Jakobshavn Glacier, bergs, especially broader ones, have capsized[12] or pack-ice allows the end to advance unduly far.[13] Others contend that projection depends upon a berg's shape; a berg which widens downwards projects, one that tapers downwards does not.[14] Bergs do apparently project in these ways as well as by tilting[15] and upward surge

by buoyancy after calving, the snout having previously been depressed as noted on the Great Karajak Glacier.[16]

Russell,[17] from his Alaskan studies, concluded that an ice-toe juts outwards, in some cases fully 300 m from the face of a glacier below sea-level and calves by buoyancy and wave-action. Tarr,[18] agreeing with Russell, pointed out that severe crevassing favoured rapid recession of the upper ice while warm waters and waves, the latter generated by berg-falls, undermined the ice about the water-line. Observations on the Muir Glacier, on the other hand, led to the diametrically opposite conclusion,[19] namely, that crevasses, undercutting by the sea, hydrostatic leverage by tides, and the concentration of subglacial drainage caused more waste below sea-level. This waste, combined with the quicker flow of the upper layers, led the latter to project forward. Crevasses facilitated the breaking off of bergs when the overhang was able by its weight to overcome the cohesion at some point behind the ice-front.

Drygalski[20] harmonised these views in an analysis which revealed three types of berg, produced at different times in different localities, two of them previously recognised in west Greenland.[21] The first are small bergs or pinnacles detached along crevasses and almost continuously from the higher parts of the face, a method stressed by Steenstrup and Heim[22] and responsible, with the numerous fractures in the snout, for the vertical glacier face. These white "false bergs" of A. E. Nordenskiöld,[23] the Eskimo's *nakarpok*[24] ("falls down from something"), are most common on crevassed glaciers which flow rapidly or the sea undercuts.[25] They vary in size from small pieces to considerable bergs and calve at all temperatures and seasons. In the second type, bergs are detached, usually in summer, under the sea and possibly from the base. They rise dripping and swaying from the water, capsize frequently, but on gaining their equilibrium fail to reach the height of the face. Since they spring from the debris-laden base[26] they are usually blue or black in contrast to the first type. A calving of this kind was witnessed in west Greenland, and more than one locality on the east coast bears the Eskimo name *Puisortok*[27] ("the place where something shoots up")—in the Merjelen See calving takes place from below.[28] The third type comprises massive bergs, split off through a glacier's entire thickness. They result from upthrust in deep water and begin with a slow rise of the snout and come to rest after calving and rhythmic oscillation. This type is rare, except in the Antarctic (its existence is denied[29]) for considerations of ice-flow and snowfall show that in Greenland at any rate bergs of these dimensions can only calve at long intervals.

An easy gradual slope of the fjord-bottom, as in front of the Jakobshavn Glacier, helps to promote large bergs, while a glacier on a steep and uneven declivity suffers many interruptions to its flow and only small pieces float away.[30]

Antarctic bergs. Bergs in the Antarctic are more uniform than in the Arctic for they have almost always the same shape and the same structure. As Captain Cook[31] was the first to notice, they are massive and flat-topped and have precipitous sides and rectangular plan (pl. VIIA, p. 192): some of them are dome-shaped,[32] possibly because they calved from surfaces which had a "roll" shape and melted more rapidly near the edge, or (in the case of those which have stranded) because of plastic flow. These "ice-islands", as Cook termed them, are more correctly speaking "snow bergs" for, unlike bergs which calve from the dissipator, they originate in the reservoir. Their

beautifully stratified névé becomes bluer and more closely stratified downwards, the intensity of the light making the grey sky behind appear distinctly reddened and the sea below dark indigo. The "black and white" and "bottle-green" bergs of the Weddell Sea quadrant owe their dark colours to morainic and wind-blown material.[33]

A berg's factor of flotation is controlled by its shape (as W. H. B. Webster[34] first observed), since draught depends on mass and not on height (see below). It is also controlled by the nature of its ice, the freight of rock-debris, and by the density of the water in which it floats. The ratio of emerged to submerged part in arctic bergs has usually been computed at 1:5 to 1:7,[35] i.e. the height above the sea, measured in feet, roughly equals the depth below, measured in fathoms. Recent investigations, conforming with the fact that sizeable bergs drift in shallow waters as over the Grand Bank of Newfoundland or across the sill of Davis Strait, suggest the following ratios[36]:

Types	*Proportions (exposed to submerged)*
Blocky, precipitous sides	1:5
Rounded	1:4
Picturesque, Greenland	1:3
Pinnacled and ridged	1:2
Last stages, horned and winged	1:1

Estimates are based upon the depth of the sea in which bergs ground, upon the dimensions and proportions of overturned bergs, and upon the density of the ice. Various figures of this have been given[37] but Bunsen's value of 0·91676[38] has generally been accepted since 1870, though recent observations[39] reduce it to 0·8997. The results differ apparently because ice is subject to internal strains and forms under different conditions and at different rates: the density in Greenland has been given as 0·911, on the Rhône Glacier as 0·880 and in North-East Land as 0·875.[40] River-ice also varies, at least temporarily, in its density with its speed of formation.[41] Discrepancies in bergs are introduced by cavities and crevasses, by aeration that may amount to 15% of the ice,[42] and by the percentage of firn which may constitute the uppermost layers even in an arctic berg. The theoretical density of pure ice derived from accurate measurements of the lattice constant of ice formed from distilled water at 0°C is 0·9168 g/cu.cm. Any deviation presumably signifies the presence of impurities including air.[43]

Observations by the International Ice Patrol (see p. 183), which show that bergs are probably more or less regular or rectangular below sea-level but tend to be pyramidal or conical above the sea, suggest a ratio of about 1:3 or 1:4.[44]

Antarctic bergs, which mainly consist of firn,[45] are less dense and therefore float proportionately higher. Their factor of flotation has been given as 1:5 though in some it is less than 1:4.[46] Unlike arctic bergs, they break away from shelf-ice which is dissected by joints due to torsional strain induced by tides. Ross Barrier in this way receded on an average 15–20 miles (*c.* 25–32 km) with a maximum of 47 miles[47] (*c.* 75 km) after Ross charted it in 1841 (provided we concur in the general view that the discrepancy between his chart and that of R. F. Scott half a century later is not due to his error in estimating distances between ship and cliff[48]). While the recession has progressed about Balloon Bight, which disappeared between 1902 and 1908,

the Barrier in general has advanced northwards between 1902 and 1911 at the rate of roughly 1600 m/annum[49] (10–16 km in all). A re-survey in 1935 showed a further advance by amounts ranging up to 22 km.[50] At present it is receding.[51]

Although the ice reaches the sea along most of the Antarctic periphery, bergs of Arctic composition and form are born in but few places, as in Graham Land (e.g. off Alexander Land, at the western end of Stefansson Strait) and off South Victoria Land. They are, however, relatively insignificant in size and remove comparatively little ice,[52] except possibly where "channel glaciers" reach the coast.

Tabular bergs, flat topped and steep sided and seemingly of Antarctic type, have been occasionally observed in the Arctic,[53] e.g. in Greenland, near White Island, and between Hope Island and Stans Foreland where they have been, in all probability, derived from North-East Land which has such bergs off its coast.[54] Others have been seen elsewhere off Spitsbergen[55] and between there and Greenland whither they may have drifted from Severnaya Zemlya.[56] Those sighted off Newfoundland have probably come from the Steenstrup and Nansen glaciers, west Greenland.[57] Tabular bergs originating in Franz Josef Land are rarely more than 25 m high.

Although Spitsbergen bergs resemble in shape and structure the Antarctic tabular bergs, those encountered elsewhere in the Arctic do so only superficially. In east Greenland, for instance, they are rafts of glacier-ice with jagged tops which may owe their shape to overturning (Nansen), as may the smooth-faced bergs of west Greenland,[58] including Kangerdlugsuak Fjord. Arctic floebergs,[59] which are flat and rectangular, also differ in origin from the typical Antarctic bergs. Winds have built up bergs, 15 m thick, off north Canada.[60]

Influence of sea-ice. The "iceberg banks" of north Greenland, in particular before the large and very productive Jakobshavn Glacier, are shoals out to sea or across the mouths of the ice-fjords upon which bergs in close array have run aground.[61] A whole fjord or bay between the barrier and the ice-face may be closely packed with bergs, cemented together as in Jakobshavn Fjord[62]—no boat has penetrated the fjord in living memory—or re-fused as a floating extension of a glacier-tongue, as before the Jungersen and Academy glaciers.[63] Of similar origin, in Drygalski's opinion,[64] are the ice-terraces of Graham Land[65] and the "confluent ice" of north Greenland.[66]

Sea-ice may deflect a glacier along a fjord's sides[67] or, like the stopper of a bottle, prevent the escape of bergs every year, as in Smith Sound and Kane Basin or in more southerly latitudes in east Greenland.[68] Exceptions occur if a glacier is strong enough to burst through the fjord-ice throughout the year.[69]

In ice-fjords, frost-bound during the winter, bergs and calf-ice remain piled up in impenetrable confusion and the entire yearly production of bergs may be carried out during the few summer months. Melting and crumbling of the winter-ice progresses from without until the narrowing strip of sea-ice can no longer retain the pent-up mass. The remnant, pushed into waves, breaks suddenly and the whole train of bergs rushes out through the fjord almost catastrophically. The train starts slowly but quickly gathers momentum and attains an hourly speed of 5–8 miles (8–13 km)—as fast as a fox can run say the Greenlanders—to the accompaniment of a deafening roar which may be audible for miles and last for days.[70] A period of sporadic bergs or a steady drift of bergs follows.

The "shooting out" (*Udskydning*[71]) is typical of west Greenland's ice-fjords. M. P. Porsild's "Torssukátak type",[72] named after the large ice-fjord of this name, is characterised by a single shooting out while his "Jakobshavn type" has numerous discharges; the type fjord's annual average is 10. The *Udskydning* becomes later polewards and as the fjords get bigger, especially if they contain islands or shoals.[73] In the highest latitudes, bergs may not escape in some years or may be permanently imprisoned,[74] as in the Antarctic or about the large and numerous glaciers in north Greenland. Thus bergs are generated on a considerable scale in favourable ice-years.

The paroxysmal emptying of ice-fjords, a phenomenon well known to the Eskimos and Danes of Greenland, is probably due, as Porsild has said, to the vast volume of water dammed up under and behind the ice at the head of the fjord during times of abnormal melting such as the foehn induces: the detachment is achieved through the agency of spring tides. Thus the shooting out in Frederikshaab Sermilik, west Greenland, is linked with the emptying of the glacier-lake Imaersartog when its waters become so deep that they lift the glacier.[75]

Débâcles are also known from the Antarctic. Thus in the Southern Ocean, 1832, 1854, 1893, 1897 and 1922 were years of abundant bergs.[76] Between 1892 and 1897, for example, the efflux of floes and bergs was so enormous that traffic between South America, Africa and Australia had to seek a more northerly track: it profoundly disturbed the monsoon régime of the Indian Ocean and caused droughts in Australia.[77] These and similar outbursts in the Arctic Ocean have been correlated with astronomical relationships,[78] which rhythmically increase and decrease the intensity of the tide-generating force, but are probably climatic and connected with the greater extent of the land-ice. The sea-ice of Weddell Sea may follow Brückner's period[79] (see p. 133) and be controlled by climatic variations.[80] The retreat of Ross Barrier (see above) has been tentatively ascribed to earth-tremors,[81] associated with faults in the Ross Sea *Senkungsfeld* which also generated the abnormal fleet of bergs the *Nimrod* met in 1908. Earthquakes have been invoked too for the unusual quantity of bergs in the Southern Seas in 1828[82] and for their exceptional size and abundance in certain years in Alaskan waters.[83]

Drygalski,[84] who discussed the periodicity of calving, concluded that with a glacier's constant movement calving should be more or less regular, though intermediate calvings, resulting from irregularities of the face or, in the Antarctic, from violent and frequent storms, might mask the intervals. Conceivably, a flood tide by deepening the water and raising the hydrostatic pressure, should favour calving. Drygalski found this influence was inconsiderable in Greenland and the time of calving independent of the state of the tide, but later experience in both the Arctic and Antarctic has demonstrated its importance.[85] It has been shown too how pieces of ice, fallen into crevasses open at low water, act as wedges at flood tide, so that a rising tide forces the ice apart.[86]

Dimensions. Heights and other dimensions of bergs, given in the early literature, as for the Antarctic,[87] are exaggerated and to be accepted with caution: it is difficult to estimate accurately from a ship or avoid over-estimating heights when observing at sea-level.

The first reliable estimates of Greenland bergs were made by Rink[88] who judged their height to average 60 m, a figure which agrees well with those afterwards obtained by Hammer, Steenstrup and others.[89] Bergs from the

Jakobshavn Glacier are higher than those from other Greenland glaciers: they have been given as 122 m (Helland), 108 m (Hammer) and 137 m (Drygalski). Of 87 Greenland bergs[90] few were more than 100 m high and the average was 70–80 m: the maximum in west Greenland was 75 m.[91] The average of the 1300 bergs seen by the International Ice Patrol in 1929 was 30 m high, 30 m wide and 120 m long.[92]

The Antarctic shelf-ice favours the calving of much bigger bergs. Twenty-one ships observed a berg 60 miles (*c.* 100 km) long in 1855.[93] *Erebus*, *Challenger*, *Aurora*, *Nimrod*, *Terra Nova*, *Endurance* and *Norwegia* all reported bergs over 20 miles (*c.* 32 km) in length,[94] as in Weddell Sea, while *Discovery II* noticed[95] in 1927, between South Orkney Islands and Graham Land, a berg 35 miles (*c.* 57 km) long (the other side, observed about the same time by another ship, was *c.* 100 miles or 160 km long) and again, in 1930, a berg 60–70 miles (*c.* 100–112 km) long between South Georgia and South Sandwich Islands. These and other huge "ice-islands" noticed after 1927 in South Georgian waters were probably derived, as the result of a tidal wave, from the unknown, south-western part of Weddell Sea where bergs are bigger than in Ross Sea. D. Mawson[96] saw a berg 40 miles (*c.* 65 km) long.

Yet these figures are exceptional: the largest bergs Ross noted were 4 miles (*c.* 6·5 km) long and the vast majority seen by the *Discovery*[97] were less than ¼ mile long (*c.* 400 m) and *c.* 37 m high. The height, usually 30–40 m, is less variable but ranges from 15 m to more than 70 m. The highest observed[98] in Ross Sea by the *Terra Nova* was *c.* 49 m, by the *François* 50 m (when tilted 88·8 m), and by the *Gauss* 50 m (and 1 km long).

Distribution. Since the Arctic ice is generally less than the land on which it rests, its bergs are relatively small and confined to the narrow ice-fjords which produce them or to the seas into which off-shore winds drift them. They are not encountered in the heart of the Arctic Ocean, though they are occasionally seen about North-East Land, around Novaya Zemlya (save in the shallow waters on the north and west), and in the neighbourhood of Franz Josef Land where their maximum thickness is 20 m or 22 m. They are rare along the Siberian coast, except near Severnaya Zemlya.

Greenland, Franz Josef Land and Novaya Zemlya are the main birthplaces of big bergs. Yet even Greenland with its numerous ice-fjords yields bergs over very limited stretches, though these are extremely productive.[99] The north coast of Peary Land[100] has only one calving glacier. The most productive are on the west north of 69° N. Lat. and on the east coast south of that latitude.[101] Thus on the west, which it is estimated gives rise annually to 7500 sizeable bergs, each 50 million cu. ft in volume,[102] the most productive glaciers are about North-east Bay and Disco Bay, viz. Sugar Loaf Bay (74° 15′ and 73° 57′), Giesecke Icefjord (73° 30′), Upernivik Icefjord (72° 53′), Rink Glacier (71° 40′), Itivdliarssuk (70° 47′), Great Karajak (70°23′), Torssukátak (70° 02′), Jakobshavn Icefjord (69° 10′) and Sermilik Icefjord (61° 13′). There is no ice-fjord of the first class between Cape Farewell and Disco Bay.

On the east, the most important are Scorseby Sound (70° 00′), Kangerdlugssuak (68° 10′)—the most productive Greenland glaciers—north Kerssuak (66° 30′), south Sermilik (66° 25′), Ikerssuak (65° 30′), Rikiutdlik (65° 00′), Igdluluarssuak (63° 40′) and Anoritôk (61° 30′).

The explanation of this striking difference in the behaviour of the two coasts lies in the iceshed's eccentricity (see p. 667) and the relation of the land to the ice-edge, and the shoals which prevent bergs from escaping on the

east coast north of 68° N. Lat.[103] Sea-ice is obstructive along both coasts, especially on the east under the influence of winds from the Icelandic Low. With the slow motion of the Humboldt Glacier it is responsible for the few bergs this glacier generates.[104]

Greenland bergs which collect in Baffin Bay, Smith Sound and the East Greenland Current—bergs from east Greenland move round Cape Farewell to a gathering ground in Davis Strait[105]—are drifted into the North Atlantic where they gravely menace shipping[106] (the "Gateway of the Icebergs", a rectangle of about 40,000 sq. miles (*c.* 100,000 sq. km), is between 47° and 43° N. east of the Grand Bank). The ice-conditions in this ocean are the subject of annual reports from the Danish Meteorological Institute summarised by C. J. H. Speerschneider (see p. 195). For the section about Newfoundland and the Grand Bank,[107] where the severity of the iceberg season is related to the pressure distribution over the North Atlantic region, reports are given by the International Ice-patrol[108]: in 1912 *c.* 1200 bergs were sighted, in 1929 over 1300 bergs, and in 1924 only 11. The number of bergs reaching positions south of the 48th Parallel averaged about 430 per annum between 1900 and 1947 and 407 between 1900 and 1953 but varied greatly[109] (fig. 39), being influenced by the preceding atmospheric pressure and temperature distribution in the North Atlantic. This patrol has been carried out for the maritime nations by the Government of the United States since 1914 (except for the war years 1940–5), following the loss of S.S. *Titanic* in 1912 (14th April) through collision with a berg.

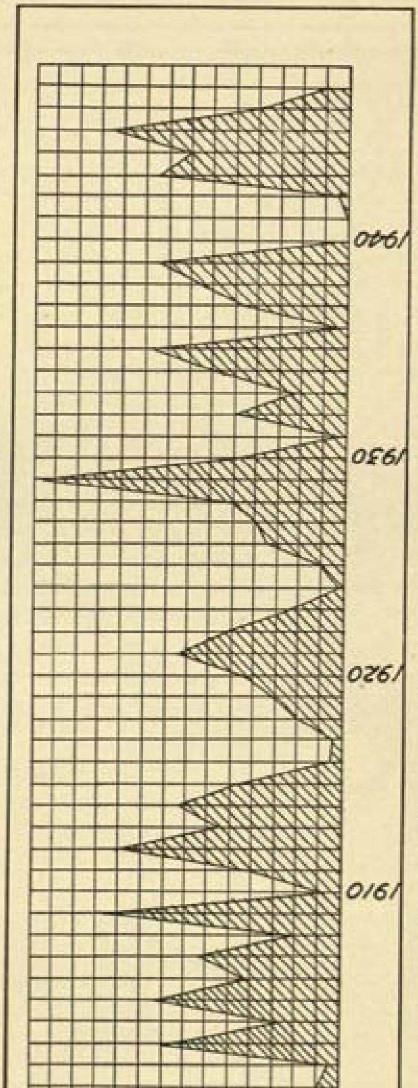


FIG. 39.—Number of icebergs south of the 48th Parallel between 1900 and 1947. United States Hydrogr. Office, *Arctic*, 2, 1949, p. 31, fig. 2.

The positions in which bergs have been sighted around the Grand Bank south of Newfoundland during the years 1900–30 are shown in the text-figure[110] (fig. 40). In ice-poor years they often remain north of 45° N. Lat. but in ice-rich years penetrate to 40° or even 30° whence they are

drifted eastwards to the neighbourhood of the Azores and into British seas (see p. 195).

Iceland and Europe liberate no bergs since their glaciers do not reach the sea[111]: the lowest lobe of the Vatnajökull descends to 60 m A.S.L. (63° 30′) and in Norway the Suphellebrae ends at 52 m. and the ice which descends to sea-level from Frostisen in Ofotenfjord in 68° 13′ N. is also a reconstructed

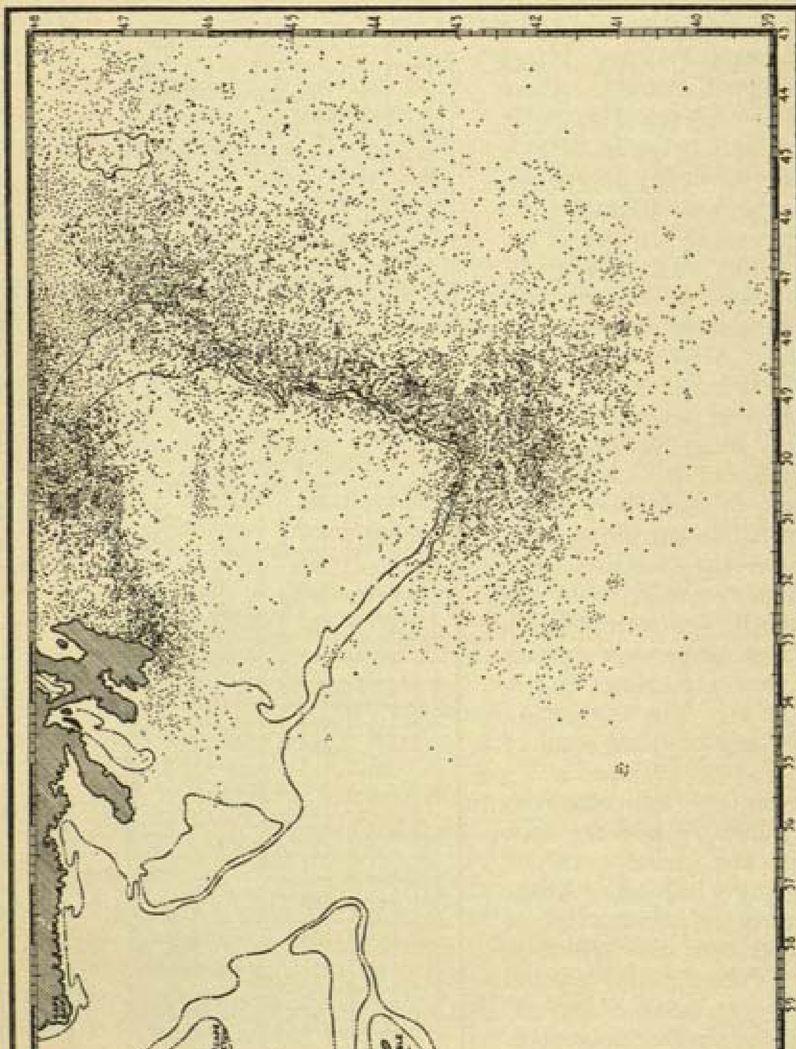


Fig. 40.—Positions of icebergs sighted over the Grand Bank between 1900 and 1930. *1577, p. 166, fig. 107.*

glacier. Spitsbergen's most productive glacier is the Negri Glacier at the head of Storfjord, though others less productive are on the east coast of North-East Land; and King James Glacier on Edge Island contributes a number of bergs annually.[112] Severnaya Zemlya is still more productive. The countless bergs from the east side of Novaya Zemlya drift to the north-north-west to the meridian of Franz Josef Land in 84° 40′ N. Lat. and then to the south-west into the Greenland Sea.[113] One small glacier on Bennett

Island yields the only bergs on the whole Siberian coast east of Cape Chelyuskin.

Ellesmere Land, alone in the Canadian Archipelago, discharges any appreciable number of bergs: the annual output on the west side of Baffin Bay and Davis Strait may be 150.[114] Bergs are produced in considerable numbers from Alaskan glaciers,[115] though floating ice is not now found in the Gulf of Alaska.

In the southern hemisphere, bergs are only calved around the Antarctic continent and subantarctic islands and in the extremity of South America where glaciers fail to reach the sea north of 46° 40′ S. Lat.—in British Columbia in North America the corresponding figure is 57° N. Lat. New Zealand glaciers, not being tidal, supply no bergs: the Franz Josef and Fox glaciers—the two longest glaciers of the west—extend to within 210 m of sea-level, while on the east, the Tasman Glacier, which is the longest and reaches to the lowest level, ends at over 600 m.[116] Antarctic bergs are often abundant in the pack-ice. Thus Captain Cook counted on 26 January, 1773, 186 from his masthead and de Gerlache saw 320 at one time in February 1898. They are, however, scarce in the range of longitude between 140° E. and 170° W., e.g. off South Victoria Land.

In both hemispheres bergs are passive. They drift into lower latitudes by the action of calving floods, by oscillations about the centre of gravity, by currents and tides, and much less by winds, unless their careers are nearly ended and they are extremely winged and pinnacled.[117] Their deep draught not infrequently takes them by undercurrents in a direction contrary to or at different rates from the floe-ice which drifts with surface winds. Projecting higher into the wind and offering more resistance, they may also move more quickly than the main pack through which they plough their way. A berg sailing through the pack with the speed of about 10 miles (16 km) a day against the wind constitutes a danger to ships which are fast in the ice.

The usual history of a west Greenland berg[118] is to be released from a fjord in summer, to reach Hudson Strait the same season, to "winter" there, and to appear off Newfoundland the following summer.

Few figures of berg production are available. Helland estimated that of the Torssukátak at 6·3 million cu. m/diem or 2·3 cu. km/annum and of the Jakobshavn Glacier at 16 million cu. m or 5·8 cu. km. Four north Greenland glaciers in 1920 may have discharged more than 150 cu. km, the product of several years calving.[119] The number calved in Greenland each year is somewhere around 10,000–15,000, divided equally between east and west,[120] while the annual wastage of the Greenland ice by calving and glacial streams is possibly one-tenth of its total mass.[121] Other arctic lands yield an annual total of about 600 bergs.[122] The annual production of icebergs in the Antarctic has been estimated at 640 cu. m.[123] Attempts have been made to derive formulae for predicting the severity of iceberg seasons based on pressures or temperatures.[124]

Weathering. Bergs are short-lived: off Greenland few are older than 2 years.[125] A berg's average volume of 50 million cu. ft (1·4 million cu. m) in Davis Strait is only 6–8 million cu. ft (*c.* 0·17–0·23 million cu. m) at the Grand Bank.[126] The height which in Disco Bay is *c.* 60 m is simultaneously halved. Bergs in their two or three years' journey from Greenland to the Grand Banks suffer a mortality of more than 90%.[127]

The wasting processes begin as soon as a berg calves. Pieces break off along joints, especially if a berg is rocked by swell. Disintegration is brought about by the high specific heat of rocks in the berg's debris-laden layers, by the action of air, noticeably in rain or mists,[128] by foehn winds as it travels along the fjord, by the hundreds of rivulets which stream off its faces, as about the Grand Bank in summer, by waves which are especially destructive in the North Atlantic and honeycomb it with caves (these are enlarged by blow holes), and by submarine melting which reduces the angles and flattens the curves—Alaskan bergs melt 10 times, west Greenland bergs 200 times, more quickly by the sea than by the atmosphere.[129] Solar radiation affects little since the sun's rays are reflected. Melting may be rapid even if the atmosphere is below 0°C[130] and is greater below water-level in winter and above it in summer.[131] This seasonal change slowly lowers a berg during the winter and raises it during the summer. Submarine calving may overturn a berg,[132] while the top layer of a tilted Antarctic berg may slide off bodily along the stratification.[133] Bergs disintegrate too by internal tension or explosive force,[134] especially at or immediately after sunrise, when the pressure of the imprisoned air (which has the same composition as ordinary air[135]) exceeds the structural strength of the ice. In north Greenland, for example, fragments the size of walnuts were discharged constantly and almost explosively and flung far away with a loud report. Eskimos are said to keep perfectly quiet when obliged to pass a berg at close quarters. The millions of tiny bubbles of compressed air, released by melting, rise by the most direct route to the surface of the sea and cause ascending currents which flute the sides of bergs.[136]

Weathering in the Antarctic enlarges the grain, diminishes the air-content, and by molecular scattering changes the colour from white to blue.[137] Arctic bergs, which are already composed of ice, alter by the release of innumerable air-bubbles to a peculiar opaque flat white, often with soft irridescent hues of green and blue.[138] Most, however, are ribboned with debris or are striped blue and green with veins of compact, transparent ice, the ancient crevasses.

Weathering sculptures bergs into innumerable shapes and into forms transitional between Arctic and Antarctic types[139]: they become pinnacled, domed, roofed or ledged. The continual surge of waves and swell developes a central bore and later a deep, wide valley, the bergs being variously styled "valley", "dry-dock", "winged", "saddle-back" or "double-horned".[140] The final stage is the "growler" which drifts nearly awash. Bergs which have spent their whole life in the pack have a more regular outline: swell, waves and sea-spray are absent and air-temperatures are more constant.[141]

If a berg's mass decreases asymmetrically, it shifts its centre of gravity and alters its trim, either by lifting the plane of flotation uniformly or by listing or overturning. Bergs which are practically as long as high or have a great height and small base are top-heavy or bottom-buoyant,[142] though those with much basal debris turn over less readily. They often capsize just after calving when their equilibrium, which depends upon a glacier's depth and degree of crevassing (this varies even in different parts of the same snout, according to the slope of the ice-fjord[143]), is apt to be unstable. Hence capsized bergs in west Greenland are confined to particular fjords, e.g. Great Karajak, while Jakobshavn and other fjords which have broad glaciers and little gradient, are almost free from them.[144] Overturning, which dense pack-ice may prevent,[145] is less prevalent in the Antarctic since the shelf-ice

from which the tabular bergs calve is regular. Sea-ice similarly floats very evenly because of its regular form and constitution.

Tilted bergs, notably in the Antarctic,[146] are recognisable by their inclined stratification and, as elsewhere, by their lines of caves and girdling incisions or notches, clearly visible miles away. These are produced by the relatively high temperatures that obtain in the surface waters. Bergs off the Grand Bank of Newfoundland have as many as three or four old water-lines inscribed across them. If the uplift is slow, the waves attack the whole of the rising surface as it emerges, if by jerks and unequal, they hollow out caves at different levels. Lines of wave-worn grooves and caves, polished on their inner side, mark successive stages in a berg's elevation where they are parallel, or a displacement of its plane of flotation if they intersect. Bergs, recently capsized, are blue in contrast to the white of sunburnt bergs[147] (see above). Rocking and capsizing of big bergs and continued breaking off of pieces induce waves which may tilt or capsize neighbouring bergs or calve new bergs.

Bergs, by their weathering, cool the air and water in their immediate vicinity. Investigations into this circulatory effect have given contradictory results[148] which suggest that it is so weak that the many temperature variations to be found in surface waters mask it. Even the most sensitive thermal recorder hardly serves as a reliable berg detector.[149] The simultaneous melting of the 1300 bergs counted in 1929 off the Grand Bank would, it was estimated, lower the temperature by 0·008°C only.[150]

Bergs provide one of the main sources of noise[151] in the "silent" North. Their calving produces a blasting sound; their break-up a booming or cannonading sound which reverberates in exaggerated tones; their overturning a swish of waves against the shore; and their weathering the note of the trickle of waters streaming off them or the drip-drip that splashes from the sides.

2. *Sea-ice*

Sea-ice in its many forms,[152] the "congealed sea" (*Mare concretum* or *Mare cronium*) of classical and later times[153] is, with river-ice, economically and socially of far greater importance than glaciers. It has been the subject of a considerable literature and of important publications,[154] in addition to the brief and incidental remarks made in innumerable books of travel throughout the centuries.[155] E. H. Smith[156] monographed the physical properties of ice ascertained by modern apparatus. Treatises have also appeared on the sea-ice of particular regions, e.g. east Greenland,[157] west Greenland,[158] Spitsbergen,[159] the Baltic,[160] Weddell Sea,[161] Ross Sea[162] and Kaiser Wilhelm II Land.[163]

Sea-ice may be classified as follows[164]:

(*a*) Primary growth types
 1. Open sea: ice-crystals; slush; ice-rind; sludge-ice; pancake-ice.
 2. Fast ice: young or bay ice; level ice; shore-ice; ice-foot (in part).

(*b*) Secondary types formed by deformation of primary types
 1. Ice-field; ice-floe; *glaçons*; growlers; cakes, lumps or fragments; brash; honeycombed ice.
 2. Hummocks; pack-ice; palaeocrystic ice.

The terms date almost entirely from W. Scoresby's classic descriptions of 1820.[165]

Mode of formation. In sea-water, with an average salinity of 35‰, the freezing point is 1·91°C and the density 1·02821, the maximum density (= 1·02822) being reached at a temperature of −3·945°C[166] as shown by T. C. Hope's classic experiments.[167] Figures of the salinities, density maxima and freezing points are set out below:

Salinity	0%	1·0%	2·0%	3·0%	3·5%
Density maximum . .	4·0°C	1·9°C	−0·3°C	−2·5°C	−3·6°C
Freezing point . .	0·0°C	−0·5°C	−1·1°C	−1·6°C	−1·9°C

Before it congeals, water contains ice as a solution carries salt. The particles of ice, as soon as they become microscopic, lack crystal form and appear as a true colloid in small discs. These flocculate and grow and build crystals clouding the water.[168]

The first outward manifestations of freezing are the obliteration of the minor ripples and the formation of frazil ice, a scum of delicate needles or scale-like crystals. About 1·25 cm, occasionally 2·5 cm in diameter, these grow very quickly in snowfall which adds greatly to the number[169] by lowering the salinity and raising the freezing point. This snow-slush (Ger. *Schneebrei*; Fr. *bouille de neige*; Russ. *snjedjura*) floats about, even at depths of several feet, and unites into rosettes, finally growing into a felt-like slush (Ger. *Eisbrei*; Fr. *graisse*; Russ. *salo*) in which the plates, oval and toothed along the edge, lie horizontally. At this early stage, the ice appears like cooling grease, grey in colour, and has little rigidity, since the mother liquor of concentrated salt solution is entangled between the crystals.

Sludge-ice (Ger. *Breieis*; Fr. *bouille de glacier*; Russ. *shuga*), after sheeting the surface, grows from below by adding vertical prisms. Under the superficial layer, the crystals stand "on edge", but being unstable, remain so only in calm weather or in sheltered leads or cracks. The wedges of plates, with brine concentrated along the planes, give sea-ice its characteristic vertical fibrous structure and its striated appearance in vertical section.[170]

The plates,[171] which are *c.* 2 cm across and shorter and thicker than in lake-ice because the sea is less calm, are sometimes horizontal, sometimes vertical; this probably depends upon the quietness of the water[172] or the strength of the diffusion currents.[173] As a rule, they are vertical, though they may be horizontal in the upper 2 in (5 cm).

The smooth film of platy ice, inconsiderably thick, grows more readily in sheltered bays, as in MacKenzie Bay, Alaska, and Melville Bay, west Greenland, to form the "bay ice" (Ger. *Jungeis*; Fr. *jeune glace*; Russ. *molokik*) of the Arctic—where the water is less saline, as off north Siberia, the freezing forms ice-rind (Ger. *Eisrinde*; Russ. *nilas, skljanka*), a thin and elastic crust. If the ice grows fairly thick and regularly, it builds the level ice (Ger. *Flacheis*; Fr. *glace plat*; Russ. *ravnyj led*). During winter the bay-ice adheres firmly to the rock or shore as the "fast ice" or "land-ice" of J. Payer and H. R. Mill, Drygalski's *Schelfeis* or Nansen's "shore-ice". Its growth is favoured by the many irregular islands of the Canadian Archipelago though the north Siberian shelf has the biggest area: in some winters it extends 270 miles (*c.* 435 km) out from the coast in the offing of the Yana River.[174] In the Baltic, freezing takes place from the coast outwards since, unlike polar regions, no old ice persists throughout the summer to act as nuclei for new ice and permit freezing to take place simultaneously over wide areas.[175]

The salinity is very variable, depending upon the season and age of the ice

and the rate of growth. In newly formed sea-ice, it is 4–5‰[176] and while not confined to the top layer, as asserted,[177] is chiefly found in that layer since the ice grows more slowly as it thickens and the salinity therefore decreases (see below): in the upper 20 cm the salinity may be 4–6‰, at a metre's depth 2–3‰. A snow cover which reduces the cooling of the surface also reduces the salinity in ice growing from below. With increasing age, the ice becomes less saline, decreasing to 1 or 2‰. Hummocky ice, as early explorers[178] including M. Frobisher and J. Davis noted, has little salinity.

When sea-water freezes, pure ice is separated from a concentrated brine which, by reason of its density, tries to soak down through the porous pulp of crystals. As this thickens, it stops the escape of the brine and retains part of the salt in small cavities and below −8·2°C also as small crystals.[179] At this time sodium sulphate and calcium carbonate also begin to separate and continue to do so with further cooling.

Salinity is a function of depth from the surface; for the more rapid the freezing, the higher is the salinity[180]: young ice, formed at an air-temperature of −40°C, has a salinity of up to 10‰, at −10°C of 4–6‰. With rapid freezing at low temperatures, the salt in the ice is deficient in chlorine, though generally the chlorine percentage of the salt in sea-ice and the brine is higher than in sea-water.[181] The salts in sea-water are selected upon freezing.[182]

The brine is mechanically trapped[183] in the interstices of the ice-plates which are themselves free from salt,[184] save for some sulphate as experiment proves. The elimination of the salt on crystallisation, which was early noticed,[185] is never more than partial unless the ice is raised above the sea, though salt drains away even if the ice is submerged. The fact that uplifted sea-ice drains until it becomes fresh enough to drink, as in the highest screwed-up ice, has been known to generations of Arctic sealers and whalers, and polar explorers have rediscovered it again and again. The salt goes into solution and drains away if the temperature rises above melting point, especially in summer, though certain salts are lost in this way[186] even if the cold is intense, e.g. −30°C. The brine, under the influence of the temperature-gradient, diffuses through the ice by a process of melting towards the warmer end and by freezing after the brine has passed.[187]

Crystalline salts or cryohydrates are exuded at the surface through the capillary cracks and form a white efflorescence of needle-shaped crystals[188] ("rassol" of the collectors of mammoth tusks on the New Siberian Islands[189]) which F. Wrangel found was more plentiful near *polynya* (see p. 192).

As the ice thickens, the actions and reactions between it and the brine become more and more complex and the law of freezing increasingly complicated. Drainage, by diminishing the salinity, reduces the plasticity; for salinity makes young ice leathery.[190] Age changes the internal properties, such as chemical composition, density, elasticity and structure—the density is very variable and ranges between 0·857 and 0·920. The ice may become granular like glacier-ice.[191]

The thickness of ice formed in a single winter[192] varies between 1·5 m and 5 m, e.g. Gulf of Bothnia, 3 m (J. C. Ross); Melville Island, 2–2·5 m (W. E. Parry); Baffin Land, *c.* 2 m (G. S. Nares); Siberian coast, *c.* 3 m (F. Wrangel); west Greenland, *c.* 3 m (I. J. Hayes) or 0·75 m (Drygalski); Franz Josef Land, 2·5 m (J. Payer); the Polar Sea, *c.* 3 m (H. U. Sverdrup); Antarctica, 1·5–4·5 m. Ice two years old[193] in the Weddell Sea was *c.* 1·5 m thick, in West Antarctica, 2·48 m. Near the ice-edge in the Arctic, the very thin surface

layer of low salinity, formed in the late summer by the melting of the old ice, favours the ice-formation of the autumn.[194]

Sea-ice grows for several years,[195] as in Melville Bay and Kane Basin, west Greenland, by snowfall, by flooding from tidal cracks, by freezing the sea and melt-water below, and by floes colliding and overriding. Growth from below has been stressed for certain areas, growth from snowfall for others[196]; the latter may explain why Antarctic sea-ice is deeper than Arctic sea-ice.[197]

Three layers of sea-ice, each becoming more complex with age, may be distinguished[198]; above, a bed of granular snow, full of air and horizontally stratified; a spongy base in process of melting; and, between the two, compact ice probably derived from fallen snow and freezing sea-water.

Growth is especially good in shallow waters which, from accessions of fresh water and precipitation, are very cold and have a low salinity. Such fast ice occurs in the flat, spacious embayments of north Siberia and on its broad continental shelf which has a mean width of 400 miles (*c.* 650 km), and in the labyrinthine channels of the Canadian Archipelago. Buffeting and frictional resistance fix the outer limit, the *sîna* of the Eskimos.

It was formerly thought that sea-ice might attain any thickness: some Antarctic bergs had indeed been regarded as heavy field-ice.[199] But sea-ice is not eternal: Nansen encountered no sea-ice in the Arctic Ocean which was 5–6 years old. Winds, poor conductivity, a freezing point lowered by brine concentrated at the base, basal disintegration, and increased pressure and melting as the ice sinks through augmenting its weight—all these set a definite limit.[200] This is a function of air and water temperatures and is probably 3·75–7 m in the Arctic.[201] North of Gaussberg it may reach 20 m.[202]

The rate of growth slows down as the limit is approached[203]; in Arctic ice it is approximately proportionate to the root of time,[204] at least initially. Malmgren[205] has graphed the annual amplitude and temperature of sea-ice at different depths obtained in the Arctic by the *Fram* and *Maud* expeditions. The temperature curve[206] is more simple in sea-ice than in bergs which are not so influenced by the sea. Waves, cold in winter, warm in summer, penetrate with a rapidly diminishing amplitude from a berg's surface to a depth of 15 m. Between the depths of 15 m and 30 m, the temperature because of the sea beneath is 1°C above the mean air temperature.

Sea-ice breaks down into *glaçons* (Ger. *Schollen*; Russ. *ldini*) or into sheets called "floe-ice" (Ger. *Flarden*; Fr. *les floes*; Russ. *oblomki*) or more extensive "field-ice" (Ger. *Eisfeld, Packeisfeld*; Fr. *champs de glace*; Russ. *ledyanoje-polje*) which is cleft by "lanes" of open water. This ice develops the following types of crack[207]: (*a*) contraction cracks induced by sudden changes of temperature; (*b*) stress or strain cracks, due to unequal loading and uneven snow accumulation (the immediate cause of the fracture is wind or swell)—they attain an appreciable width as soon as made and their sides float at different levels; (*c*) pressure cracks, formed when young and plastic ice is bent or ice is rafted by blizzards or swells from the open sea; shock or concussion cracks, opened particularly in ice which is in a state of tension; and (*d*) torsion cracks. Tidal cracks, single or multiple, especially in confined bays with gently shelving shores, constitute a special case of strain crack which follows the shore closely; in the midst of the pack, the rising tide rends the ice and expels the air imprisoned below the ice so that it escapes with the noise of thunder[208] (pl. VIIIA, facing p. 193).

Floes travel at speeds which depend upon winds, currents, tides, draught and

the nature of the floe surface; hummocks act as sails. Turning frequently produces a "screwing pack". Adjacent floes dovetail into or override one another, and by piling up and telescoping build pressure ridges. These sink to the position of equilibrium their density requires and have huge cracks roughly parallel with their sides (pl. VIIIA, facing p. 193).

Floes are curled and twisted along their edges and forced up into pressure ridges or *skrugarer* of brecciated ice, the north Siberian *toros*[209] (plur. *torosi*; Engl. toroses). In this, granular ice is substituted for platy structure.[210] The whole field is a chaos of "hummocky ice" in which the hummocky floes are wholly or partially recemented together and the thinner parts or frozen leads are buckled into overfolds and allied pressure structures. "Laminated ice" results if thin sheets are telescoped. Pressure ridges, which peculiarly characterise the margins of the Arctic basin, are rarely more than 6–8 m or 18 m high[211]: the ice through which the *Fram* drifted was seldom 9 m thick,[212] though ramparts, even 30 m high, have been observed north of Greenland.[213] Pressure ridges may arise[214] from screwing, from winds, currents and tides, or from bergs passing through young ice. Even shelf-ice may be puckered in front by the impact of other ice.[215]

"Paleocrystic ice",[216] first found by Nares[217] in Kane Basin, is a singularly chaotic, hummocky pack, scarcely distinguishable sometimes from glacier-ice, the undulating surface being blue and miles in extent. Its thickness is considerable (7–8 m) and its age at least 25 years. It also occurs north of Greenland (Robeson Channel), Grinnell Land and Grant Land, especially south of Peary's Big Lead and where the polar ice is forced into the funnel-like opening into Baffin Bay,[218] as well as along the east coast of Novaya Zemlya and around Franz Josef Land (Fridtjof Nansen Land). Of somewhat obscure origin, it has been thought to have been derived from bergs (Nares) or from *Sikussak* (Peary), and in the case of the region north of Asia from the fresh water of the great rivers of that continent which froze and so did not mix with the underlying saline waters. Pressure and heavy brecciation probably produced it from old floes, the rugged projections being rounded off by melting (Nansen) where currents, winds and the sea's configuration cause congestion in the Arctic Ocean. Sea-ice has in this way become rough and granular like glaciers[219] to form the oldest fast ice known. The *Sikussak* of the Eskimos,[220] limited geographically to the calm fjords of north Greenland, probably arose by covering the fjord-ice, of many years' duration, with snow and by melting the underside, so that ice formed from snow gradually replaced the original ice.[221]

Young ice, subject to swell, divides into countless hexagonal, subangular or roughly circular discs up to 3–5 cm thick and to 2 or 3 m across.[222] These cakes of "pancake ice" (Ger. *Pfannkucheneis*; *Drehschollenesis*; Fr. *omelates de glace*; Russ. *blinchatij led*) jostle each other continually and become round, with up-turned rims.[223] Several may unite into a larger floe or compound pancake-ice, likewise rounded into pancake form. Although pancake-ice may result if old sea-ice breaks up by rotation and friction,[224] it is more commonly an early stage in the growth of sea-ice[225] when the sea freezes while rippled by wind or disturbed by swell, as illustrated on Lake Geneva and Lake Baikal. "Ball ice", of a diameter of 2·5–5·0 cm, which has been observed in "streams" in Antarctic waters, may arise possibly from the coalescing of frazil ice particles and their subsequent rounding by collision or wave action, from small pancakes by the

same process, or from an agglomeration of snow-flakes rounded by water-movements.[226]

The unbroken ice of winter disintegrates in spring into pack-ice.[227] The floes become increasingly loose, less heavy, smaller and rounded. When not in contact, they form "open pack", when pressed together, "close pack". Nansen[228] has well described the mobility and constant regrouping of the floes. The pack drifts with the shifting winds and tides,[229] and because it is so vast may have contrary winds in different parts. Convergent winds crush it, as noticed already, into hummocks and pressure-ridges, while divergent winds open it along cracks or fissures which may widen into navigable "wakes", "lanes" or "leads" (Ger. *Wasserstrasse*; Russ. *razvodje*). Shore-water (Ger. *Küstenwasser*; Russ.' *vodyanoi zabereg*) often develops along the coast in summer, especially near the New Siberian Islands and in places in east Greenland. Offshore winds broaden the pack-ice stream and scatter its outer floes while on-shore winds narrow it and drive it against the land.

While snow-water or pools (Ger. *Schneewasser*; Russ. *Snezhnitsa*) abound on the ice, larger open water spaces, the *polynya* of Russian explorers, which are of various shapes, tend to occupy definite positions. Thus the Great Siberian polynya[230] which stretches, with interruptions, for several hundred miles westwards from the New Siberian Islands and is formed by the pack-ice drifting upon them, is controlled by the isobars—a north wind may close it or cover it with young ice. The North Water at the head of Baffin Bay and near the entrance to Lancaster, Jones and Smith Sounds has excited the interest and curiosity of explorers for two centuries. It has been attributed to the emergence of relatively warm water from the south[231] (an offshoot of the Gulf Stream). More probably it results from the strong fast ice in Smith Sound which resists the current and from the sweeping away of the ice on the south to leave open water behind it[232]—the low pressure above the polynya induces inflowing winds which in turn cause a rough sea that constantly acts upon any new ice and so keeps the sea open.[233]

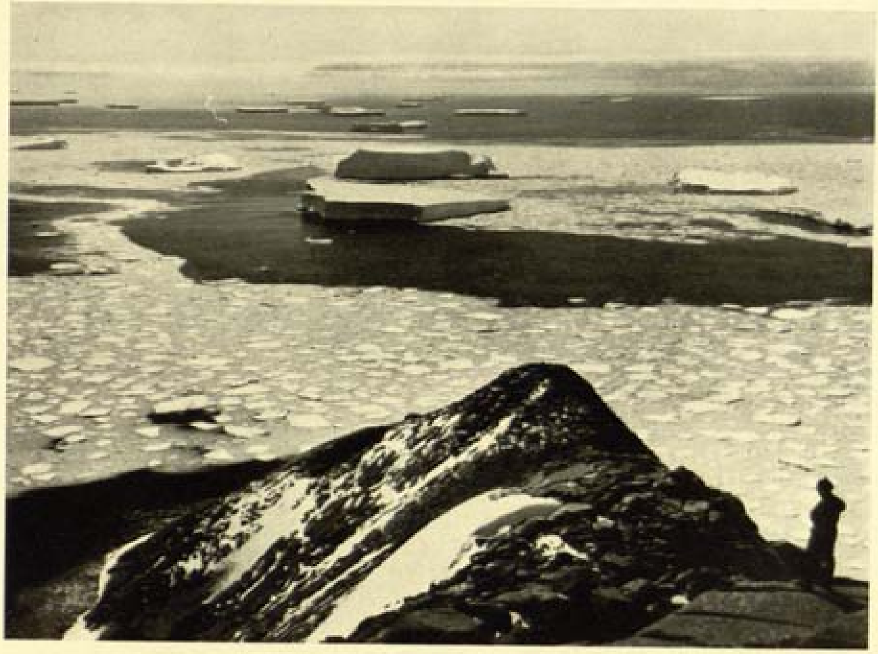
Peary's Big Lead,[234] 300 miles (*c.* 480 km) long and seldom more than 1 mile (1·6 km) wide, coincides with the continental edge along the 84th Parallel from Grant Land to Greenland (40°–60° W. Long.) and with the line of shearing of the polar pack past the palaeocrystic ice.

Open water also occurs in several places on the east coast of Greenland[235] where it is associated with bird cliffs and with former Eskimo settlements. It is due primarily to strong heating of the air over the ice-free coast land and a considerable flow of thaw-water out of the fjords.

Lanes are recognisible from a distance by the dark streaks in the clouds styled "water sky" (Ger. *Wasserschatten*). This differs strikingly from "ice-blink" (Dan. *isblink*), a diffuse reflection or narrow band of light of peculiar whiteness which ice throws upon the clouds. The first term was introduced by J. R. Forster, the second by H. E. Parry.[236]

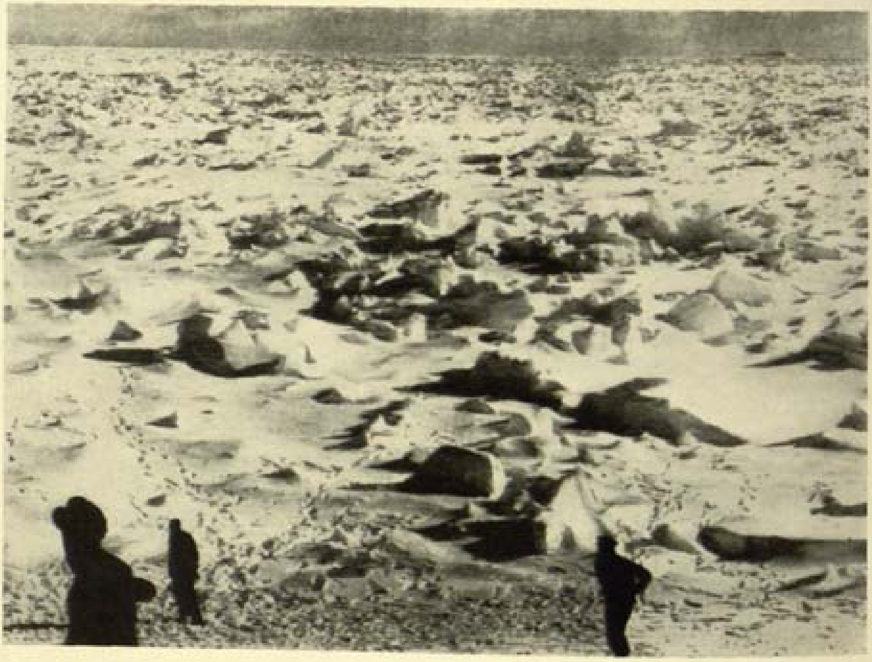
The ice-edge depends for its appearance upon the wind.[237] Winds parallel with the edge form tongues, especially at the bends of currents, and those blowing away from the edge tear off the tongues into strips or streams or belts.

Sea-ice dissolves under the influence of sea, mist and rain, though currents also play a part as is shown by the countless diatoms discovered between the basal plates of fjord-ice in west Greenland[238]—the diatoms and other organisms also contribute to the decay. Weyprecht found a loss of 58% in



A. Antarctic icebergs aground in 120 fathoms north of Cape Batterbee, Enderby Land [B.A.N.Z. Expedition 1929-1931]

B. Ice-foot near Cape Calhoun, Greenland [Royal Geographical Society]



A. Hummocked ice, Weddell Sea, Antarctic [E. A. A. Shackleton]

B. Floebergs with leads, East Greenland, Antarctic
[Scott Polar Research Institute ; photo by F. Spencer Chapman]

$7\frac{1}{2}$ months. Fast ice begins to melt near the shore under the action of melt-water and of foreign matter from the land. The hummocky ice becomes moutonnée. It is naturally the last to survive and constitutes the growlers or "bergy bits" (Russ. *nesyak*) or floebergs. Fantastically honeycombed "rotten ice" or "brash ice", fragmented by storms, fringes the extreme edge of the pack and accumulates between the floes or along the shore.

Distribution. Drift-ice occurs in all polar seas—25 years' observations for the Baltic were summarised in 1930[239]—and extends over 22·9% of the total area of the oceans.[240] In the Antarctic, it comprises both berg and sea-ice, in the Arctic, sea-ice alone or in great preponderance. In the Davis Strait—Baffin Bay region, for example, bergs are only about 2% of the volume of the sea-ice.[241] The latter is most extensive in enclosed seas and least extensive in open, stormy seas, such as the Southern Ocean.[242] Transehe,[243] who classified sea-ice genetically, agreed with A. Kolchak[244] in differentiating between Arctic pack or old rafted ice, situated mainly in the heart of the Arctic Ocean, and pack-ice found chiefly in the outer zone.

The drift-ice of the Antarctic differs from that of the Arctic not in greater thickness, but in the absence of a coastal land lane (except on the west side of Graham Land) and of the heavy piled-up pack-ice or wild toroses; the ice is less brittle, has more opportunity to spread freely, and drifts in waves which are more destructive and in currents which are less liable to sudden changes in trend. It encircles the continent in a broad belt, roughly concentric with, and except in summer, fitting closely to the continental border and usually well within the Antarctic Convergence.[245] Its breadth, as encountered for example by expeditions in Ross Sea,[246] varies with the season[247] and from year to year according to the weather of the previous season and the distribution of the air-pressure.[248] In summer, particularly in late February or early March, it is least and, except east of Ross Sea and in Weddell Sea, forms a narrow band along the coast. The farthest northern limit is reached in late winter and spring, the edge lying in much the same position from July to October. The seasonal range varies considerably and is greater in the Atlantic sector and probably least in the Bellingshausen Sea and near Adélie Land. With the exception of the Pacific sector and probably the east coast of Graham Land, nearly all parts of the fixed shelf-ice or continental coast are probably free from pack-ice from time to time during the later summer. The ice-edge is not even but follows a tortuous course. The ice-zone in Ross Sea is most readily traversed between 175° and 180° E. Long. in contrast to Weddell Sea which is most beset with ice in the western half of the sea (see below).

The limits, depicted on several published maps[249] (fig. 41), show that while sea-ice occupies infinitely more surface than shelf-ice, bergs,[250] which unlike sea-ice are distributed by currents and not by winds, drift over 23·7% of the whole southern hemisphere[251] (see below) to near New Zealand (55° S. Lat.), Tasmania, South Africa, and Tristan da Cunha, or 10–15° farther north than pack-ice, though they are rarely reported north of 35° S. in the Atlantic Ocean, north of 45° S. in the Indian Ocean, or north of 50° S. in the Pacific Ocean. They reach their lowest latitude in 26° 30′ S. in the Atlantic Ocean[252] because of the submarine ridge extending from Graham Land to South Sandwich Islands and of the cold Falkland Current and the Bouvet Current. The latter moves under the influence of the westerly winds east-north-east from the west side of Weddell Sea in 63° S. Lat. and even into

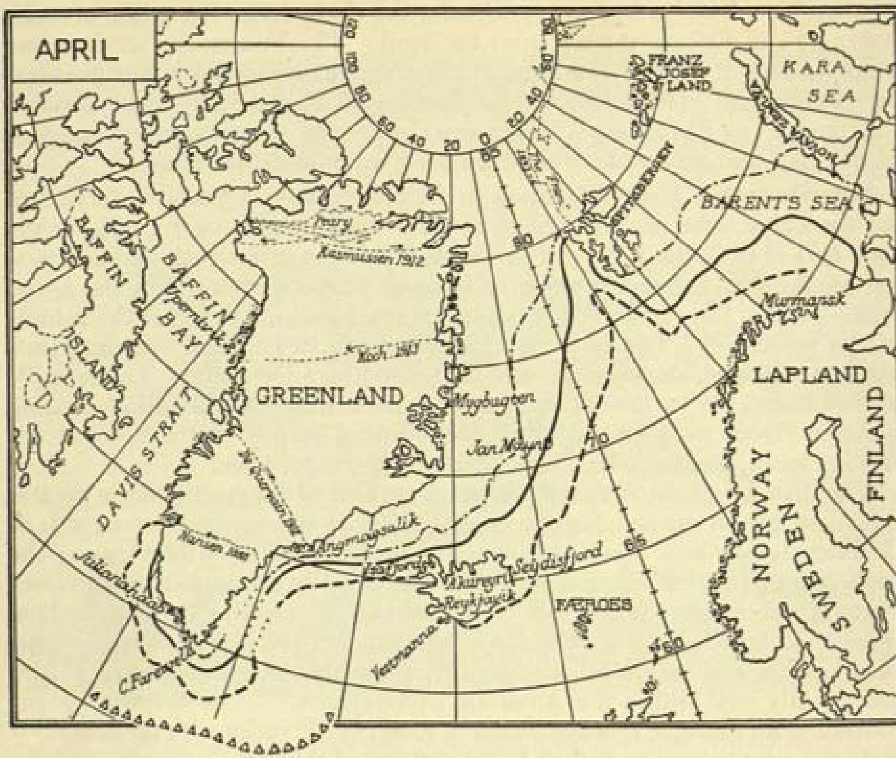


FIG. 41.—Average limits (full line) and extreme limits (broken lines) of the drift-ice in April in the North Atlantic Ocean. Extreme limit of icebergs is shown by a chain of triangles. Danish Meteor. Inst., *1551*, II, p. 16, fig. 4.

the Indian Ocean; it influences the isotherms and the limits of the drift-ice,[253] leaving an ice-free zone and a warmer area to the south of it. The different behaviour of Ross Sea in this as in other respects is due to the shape and extent of the inlet and to the fact that the cold current of its west coast, being still under the control of the easterly winds, is turned to the north-west at Cape Adare in 71° S. This sea, therefore, has little influence on the sea-ice beyond its immediate neighbourhood. Its pack-ice probably moves only half as quickly as that in Weddell Sea.[254]

The limit of the sea-ice, which like winds, currents and isotherms, etc., conforms more with the parallels of latitude than it does in the Arctic, is well defined: it is related to the East Wind whose northern boundary is approximately 65° S. and to the south of Australia lies near the Antarctic coast. Its nearness is betrayed by various signs,[255] e.g. the Antarctic Petrel (within 400 miles or 640 km), the Snow Petrel (within 100 miles or 160 km), and the Antarctic Tern (at the ice-edge), by the ice-blink, by the sudden decrease in the swell and by an abrupt drop in the sea-temperature of about 1·0–1·5°C which usually occurs 10–20 miles (16–32 km) north of the ice-edge. It encloses an area (mean pack-ice) of 22,610,000 sq. km.[256] The limit exhibits four shallow but wide "bays",[257] namely, south of New Zealand, south of Cape Horn, west of Bouvet, and south-east of Kerguelen; these are due to the position and direction of cold and warm currents.

Arctic limits, though less regular, are better defined[258] since there is less

scattering. They vary much from season to season, e.g. during the summer months of the Barents Sea region[259] (fig. 42), and from year to year, in accord with the winds and currents of the previous winters.[260] Bergs do not attain the low latitudes they do in the Antarctic because they are much smaller, start from higher latitudes, and drift in warmer seas and currents. They are absent from coastal waters in Europe and Asia and from the North Pacific but extend to 30° N. Lat. in the Atlantic Ocean.[261] Remnants, caught up in oceanic vortices, have been discovered in British seas[262] (see p. 184), as off the Orkney Islands and west of Mull and in the English Channel, and once off the Murman coast and North Cape, Norway,[263] and even near Bermuda.[264] Yet a list of all the extraordinary berg drifts showed that in 20 years only 24 bergs passed the 40th Parallel.[265]

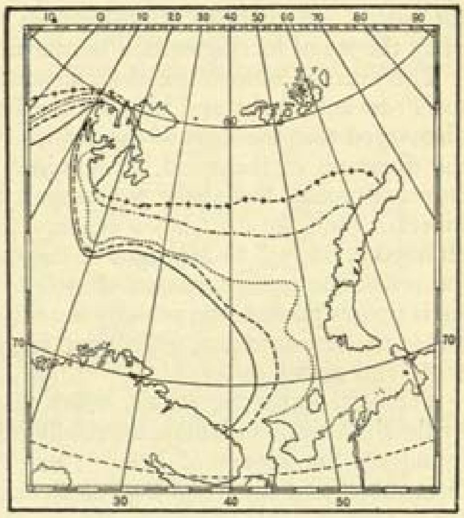


FIG. 42.

——— April
– – – – May
······ June
–·–·–· July
–+–+–+ August

Map of the mean ice-conditions in each of the summer months in the Barents Sea. B. Schulz, *J. Conseil* 5, 1930, p. 301, fig. 6.

The influence of the Gulf Stream is seen in the pressing back of the icebergs to about Bear Island and of the pack-ice to the north-west of Spitsbergen ("Whaler Bay") in Lat. 81° N. and the "North Bay" (*Nord-Bukta*) found especially in June in 72–75° N. Lat., 200 km north of Jan Mayen.[266] Generally, the waters of Iceland, Jan Mayen and West Spitsbergen are ice-free, though in early spring in some years the limit of sea-ice is near north-west Iceland, touches Jan Mayen and closes the west of Spitsbergen and Bear Island.

The limits and state of the Arctic drift-ice are given in various publications, including the Ice Atlas of the Northern Hemisphere, Hydrographic Office, U.S. Navy, No. 550, 1946 (which gives the distribution of sea-ice and river-ice of the hemisphere, and the distribution of sea-ice in the Grand Bank region, Baltic Sea, Black Sea, White Sea and Sea of Okhotsk) and the Annual Reports of the State of the Ice in the Arctic Seas (*Isforholdene i de arktiske Have*), published annually between 1894 and 1939 and since 1947 by the Danish Meteorological Office, Copenhagen.[267] In these a general summary is followed by detailed information on (1) the waters around Novaya Zemlya and Spitsbergen, (2) Greenland Sea and Denmark Strait, (3) North Atlantic, (4) Davis Strait, Baffin Bay, Hudson Bay and Strait, and (5) Siberian Sea, Bering Sea and Strait, and Beaufort Sea. Reports of the state of the ice in Davis Strait[268] (1820–1930) and in the seas off the U.S.S.R. for the winters 1924–31,[269] including the Kara Sea, have also been published. In winter, ice forms over the whole of the wide continental shelf of Siberia, because this is shallow and because the large Siberian rivers dilute its sea-water: the thickness in late summer averages *c.* 2 m, and in late winter 3·5 m.

The existence of an open polar sea within the pack-ice was often discussed in the 19th century.[270] It is now known, however, that the deep centre of the Arctic Basin, *c.* 2 million sq. miles (*c.* 5 million sq. km) in extent or 70% of the whole basin, is permanently occupied by old and heavy ice, the "polar ice-cap"[271] ("North Polar pack", "polar pack-ice", "central ice", "North Pole Ice"). Characterised by great solidity, vast fields and massive rafted hummocks, this ice is centred about the Ice Pole or Pole of Inaccessibility (83–85° N. Lat., 170–180° W. Long)—the eccentricity of this pole is due to the great "bays" on the Atlantic side (see below). Together with the Antarctic ice-sheet, this ice constitutes to-day the greatest *terra incognita* of the earth. Its margins fluctuate with the seasons, and cold ocean currents carry the ice of its fragmented border to warmer climes.

The average ratio of ice-drift to surface wind velocities[272] was found for the *Fram* and *Sedov* and later for the Beaufort Sea to be 1–2%. The *Fram* discovered that the drift of the sea-ice in the Arctic Ocean was not at 45° to the direction of the wind, as required by V. W. Ekman's theory of wind currents (1902), but about 30° to the right of that direction. On the north Siberian shelf the drift has a velocity of 1·75% that of a given wind, and is directed at 28–33° to the right of the wind—the discrepancy may be due to the resistance against motion offered by the ice itself. In spring, when the ice is tightly packed, the velocity is 1·4% and the angle of deviation about 15°, but in the late summer, when open lanes exist, the corresponding figures are 2·4% and about 40°.

The drift across the Arctic, which takes 4–5 years to accomplish, is shown by the drift-wood (conifers, especially Siberian larch) from the Siberian rivers found on the shores of Franz Josef Land, Spitsbergen, Jan Mayen and Greenland, and that from the Mackenzie River which is found to the west as far as Alaska. It is also proved by the drift of the *Fram*[273] and the *Jeanette*[274] and of the Russian North Polar station of 1937[275] which showed a daily average of 9·1 km that increased southwards and became as high as 42 km, and by the drift of the *Maud, Sedov, Lenin* and *St. Anna*.[276] It is depicted on the accompanying map[277] (fig. 43). The strength of the East Greenland Current, which consists of polar water in the main north of the Icelandic-Greenland Ridge (derived from cold surface waters of the polar basin) and of Atlantic water south of the ridge,[278] is shown by the fact that the bulk of the *c.* 5000 cu. km of river-water, the *c.* 30,000 cu. km of Pacific water, and the more than 100,000 cu. km of warm Atlantic water escapes by this route.[279]

The pack-ice (Ger. *Packeis*; Fr. *glace de pack*; Russ. *pak*), whose general movement is anticyclonic, invades the North Atlantic Ocean by two routes: (*a*) along east Greenland (see above and below) and (*b*) along east Canada. This stream is fed into Baffin Bay and Davis Strait through the tortuous channels of the Arctic Archipelago. There is no discharge through the shallow Bering Strait, though much of the northern part of the sea freezes over during the winter.[280]

The limits of the Arctic pack move with the shifting of the Arctic wind-shed which forces the polar ice to approach now the Pacific, now the Atlantic side. A 4–5-year period is recognisable near Iceland, as the inhabitants of that island believe, and has been proved for the sea between Novaya Zemlya and Cape Farewell and Jan Mayen.[281] It seems to exist also on the opposite side of the Arctic Ocean though the evidence here is less full and precise.[282] The temperature of the Gulf Stream not only influences the number of bergs

reported south of the 48th Parallel in the west Atlantic but the snow-cover in Barents Sea.[283] An unusually warm period allowed Franz Josef Land to be circumnavigated on the north in 1932[284] and open water to extend in 1935 to 82° 41′, the highest point ever reached freely by a ship.

The mutual influence of pack-ice and air-pressure in their distribution has been shown for the northern and southern hemispheres.[285] Cyclone tracks, for example, tend to swing equatorwards in years of heavy ice.

The drift-ice of east Greenland,[286] which is narrowest in August and September and a few hundred kilometres broad in winter, encloses Jan Mayen, and reaches about the north coast of Iceland whence currents carry it in late winter or early spring along the east coast. It moves with a considerable velocity[287] and consists of three elements: (1) an outer fringe of flat, young floes averaging 1–1·5 m thick and derived from seas

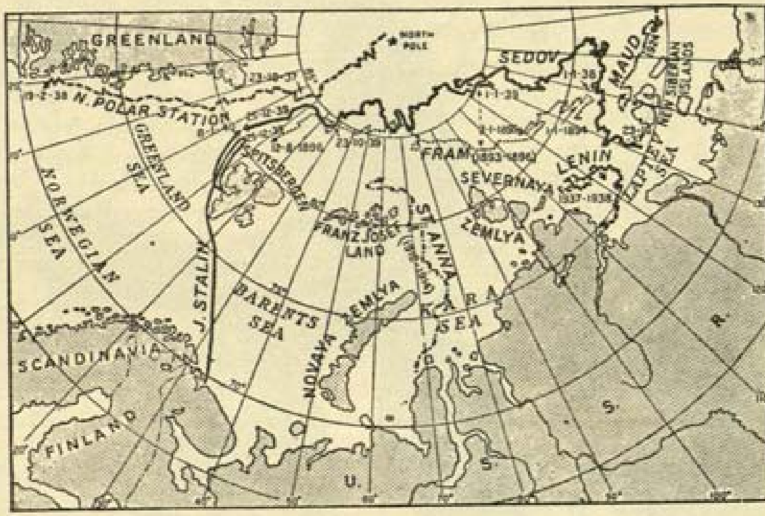


FIG. 43.—Drifts of the *Maud*, *St. Anna*, *Fram*, *Lenin*, *Sedov*, *Joseph Stalin* and Papanin's North Polar Expedition in the Arctic Ocean. N. N. Zubov, *N.* 145, 1940, p. 537; cf. C. J. Webster, *Arctic*, 7, 1954, p. 61, fig. 2.

north of Spitsbergen—in a "close season" (Ger. *Südeisjahr*) the east Greenland and east Spitsbergen ice unite; (2) heavy polar ice, the *Storis* of the Danes—this is a translation of the Eskimo word *sikorssuit* (great ice)—3–5 years old and 3 m thick which is subject to pulsations in the polar basin,[288] especially in winters; (3) and a wide inner zone of floes, some of them hundreds of square miles in area, descended from fast ice (see above).

The ice along this coast is so thick and dense that it is impassable for ships from north Greenland to 77° N. Lat. Between the parallels of 77° and 70° N. it broadens out and becomes passable and is skirted by a coastal channel kept open by melt-waters. From 70° N. to Cape Farewell, the coast is again blocked by ice of the Irminger Current, though at the cape the quantity is much reduced through melting, rain and sea. West of Cape Farewell, south-east winds, blowing parallel to the coast, keep the pack open so that Ivigut and Julianehaab, where the *Storis* makes its appearance on an average about

25 January, are accessible to ships all the year round except occasionally when south-west winds prevail.[289]

The distribution of the various types of ice off north Greenland was mapped by L. Koch[290] and in Davis Strait by Speerschneider[291] (1820–1920).

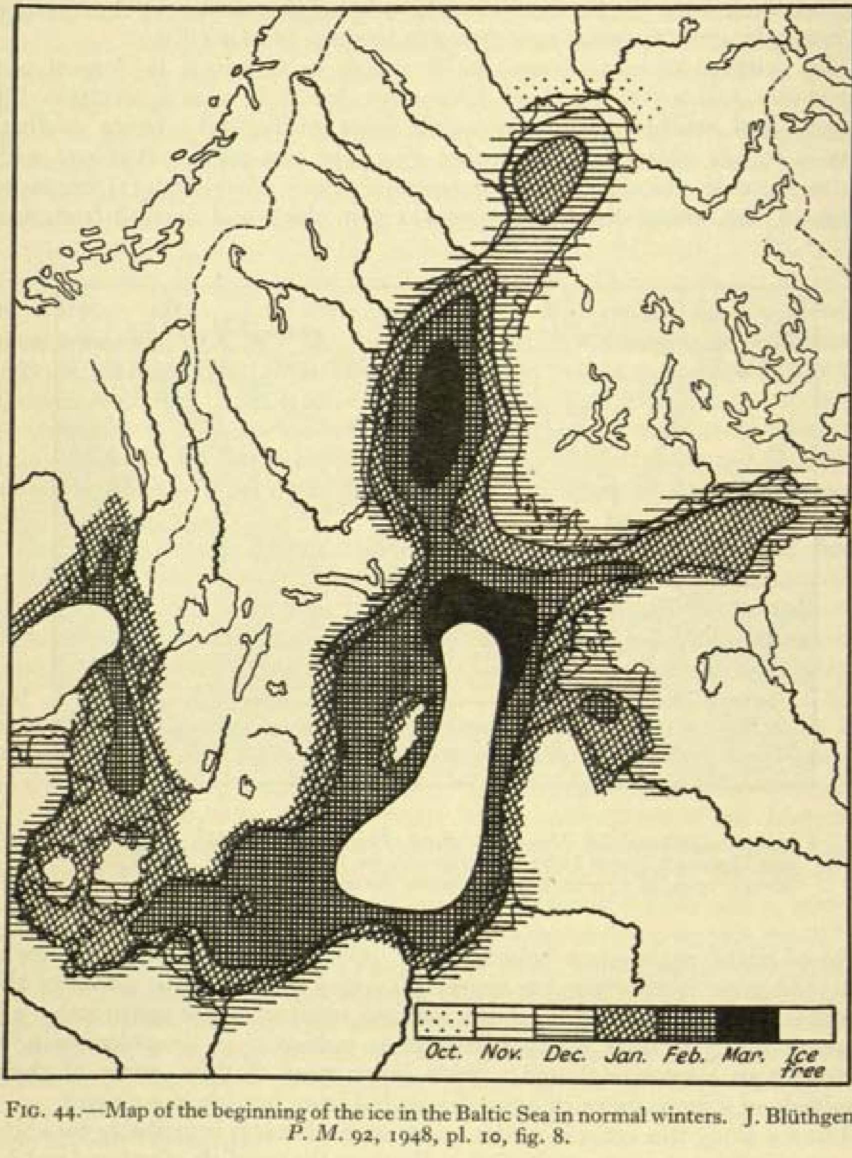


FIG. 44.—Map of the beginning of the ice in the Baltic Sea in normal winters. J. Blüthgen, *P. M.* 92, 1948, pl. 10, fig. 8.

In this strait and Baffin Bay, the winter ice consists of fast ice along the shores, and in the fjords of Baffin Land and of Greenland north of Holsteinborg, of central pack-ice from Lancaster, Jones and Smith Sounds, and of *Storis* or heavy arctic ice from the East Greenland Current. The yearly reports of the Ice Patrol (see p. 183) give much information of the ice-conditions of the North Atlantic,[292] particularly of the drift-ice in the neighbourhood of the

Grand Bank which consists of sea-ice from Labrador and berg or *Westeis* from west Greenland. The *Atlas der Eisverhältnisse im Nordatlantischen Ozean* (*Deutsche Seewarte*), 1944 gives the average ice-conditions for this ocean during the years 1919–43, including the winter months. A later edition was issued in 1950 (J. Büdel). The conditions in Finnish and Russian waters are given in the *Atlas der Vereisungsverhältnisse der Küstengewässer Russlands und Finnlands* (*Deutsche Seewarte*), 1942.

The amount of floating ice, melted per annum, has been calculated[293] to be 32 million cu. km for the southern hemisphere and 7 million for the Arctic Ocean, while the difference in range between winter and summer ice-fields has been estimated at one-eighth of the earth's surface.[294] According to Krummel,[295] 12,700 cu. km of drift-ice leave the central Arctic basin annually between Greenland and Spitsbergen, 5000 by Baffin Bay, and 2000 between Bear Island and Franz Josef Land.

In the Baltic,[296] the parts most strongly affected are the gulfs of Bothnia, Finland and Riga—the conditions in the Barents Sea greatly influence those in the Gulf of Bothnia. Fast ice occurs along the coast and between the islands: the average thickness in the Gulf of Bothnia is 70–75 cm. Immediately outside this zone is a narrow band of broken but firm ice followed outside by pack-ice and polynyas and loose and drifting sea-ice which covers large areas and sometimes assumes the shape of pancake ice. In severer and longer winters, the duration of the ice along the coast and among the islands departs much less from normal than at the outer limit of the ice. The average duration of the ice (in weeks) in the Gulf of Bothnia and during very severe and during very mild winters has been investigated and mapped,[297] also the *isokryonen* for the mean beginning and ending of the ice-period in the Baltic[298] (fig. 44). Fast ice appears in the autumn in shallow and protected coastal bays on the northern part of the Gulf of Bothnia and inner angle of Gulf of Finland. Thence it extends slowly outwards though with regressions due to temporary ameliorations of conditions. In the southern Baltic, the duration is longest in the shallower parts and is controlled by the severity of the frost. In the open Baltic, the formation of ice is hindered by convection currents which bring up deeper waters.

Hudson Bay,[299] which is open to the influence of both Arctic and Atlantic waters, freezes over in the late fall and is completely frozen from January until June, except for a coastal lead separating fast ice from the central pack and small temporary leads or fissures in the latter which are produced by storms and tides and other stresses. The drift is counter-clockwise: the ice-presses against the east side of the bay and finally escapes northwards into Hudson Strait.

3. *Lake- and River-Ice*

Detailed studies have been made of the ice and ice-conditions of many lakes, the most important of which are Lake Balaton,[300] Lake Lunz,[301] lakes in the Austrian and eastern Alps,[302] Switzerland,[303] Bavaria,[304] Scandinavia[305] and Poland,[306] and the Caspian Sea[307] where the shallowness of the sea and its accessions of fresh water in the northern part favour the formation of ice—the mean duration in the north is 90 days.

The time when lakes freeze depends upon thermal and hydrodynamic conditions,[308] viz. the temperature of the air, the strength of the wind, the intensity of the radiation from the water and the heat supply of the lakes on the

eve of freezing—this is one of the main reasons why lakes freeze at different times. It is also affected by the lake's depth, the character of the shore, the presence of islands and bays and the strength of the water-currents.[309]

Equiglacial lines or "congelonts" were subdivided by Rijkatschew[310] into *isotaks* or lines of synchronous freezing, *isopectics* or lines of synchronous thaw and *isopagi* or lines of equal ice-duration. He constructed such lines for Europe and Asiatic Russia and found that the curves were directed southwards in accord with the isotherm of 0°C. The *isotaks* and *isopectics* of European Russia are given in the following map[311] (fig. 45). Such lines have also been examined for Fennoscandia,[312] the Baltic Sea,[313] north Bohemia,[314] the Danube Basin,[315] Galicia,[316] the Caspian Sea[317] and the lakes of north Asia.[318] G. Schwalbe[319] investigated the ice-conditions on German rivers and found the extreme north-east had on an average 104 ice-days during the year with 86 days of "fast ice". The number of such days gradually diminished across the Vistula, Oder, Elbe and Weser to the Rhine which had only 16 and 13 days respectively and the Ems 10 and 5·1 days. The number in Poland increased eastwards: it was 61–70 on the Warthe and Vistula (at Warsaw), 71–80 on the Niemen, 81–90 on the Dniester, and 111–120 on the Prypeć. Congelonts are related to the mean winter temperature, modified by such other factors as the river's velocity and the condition of the tributaries. Observations prove that sluggish rivers with low fall freeze earlier and are freed later than steep and rapid rivers.[320]

The duration in days of the ice-cover on mountain lakes rises with altitude, e.g. Upper Arose Lake (1738 m), 150–160; Great St. Bernard Lake (2446 m), 211–230; Lej Sgrischus (2640 m), 240–265 and Lej della Pischa (2780 m), 365.[321]

The first ice, particularly in a flowing stream or on a broad lake fully exposed to a strong wind, is usually the frazil- or anchor-ice (which forms in turbulent waters), together with snow- or ice-scum. This floats and gradually builds big fields which quickly "set" into a solid, frozen sheet which grows continuously by conduction.[322] The pans of ice are carried downstream and touch the cold shore where they freeze tight, catch other pancakes and finally jam the river from shore to shore. Pancake-ice also forms even in temperate latitudes,[323] as in France and Switzerland and on the Pennine Chain of England.

Once a continuous cover is created the freezing process enters a new phase. Heat is lost only by conduction, and since ice is a poor conductor the temperature gradient in the ice is steep.

In all kinds of water-ice, e.g. stream-, lake-, fjord- or sea-ice, as well as in encrusting ice, ice-stalactites and ice-stalagmites,[324] plates constitute the fundamental ice-form (prisms are plates united). They owe their formation to intermittent crystallisation.[325] Each plate is uni-axial, as was shown by Brewster, Klocke, Reusch and Tyndall, and elasticity experiments and pressure-figures have confirmed.[326] The axes of the crystals, often described,[327] are parallel to the direction of the temperature-gradient or normal to the water-surface as seen in polarised light.[328] During growth, plates are added to the base of the prisms; the purer the water the greater is the length which may be several inches.[329]

The prisms' orientation perpendicular to the cooling surface, seen too in their radial arrangement in dust-wells (see p. 60) and their horizontal disposition in crevasses (see p. 49), may be because heat is more readily conducted

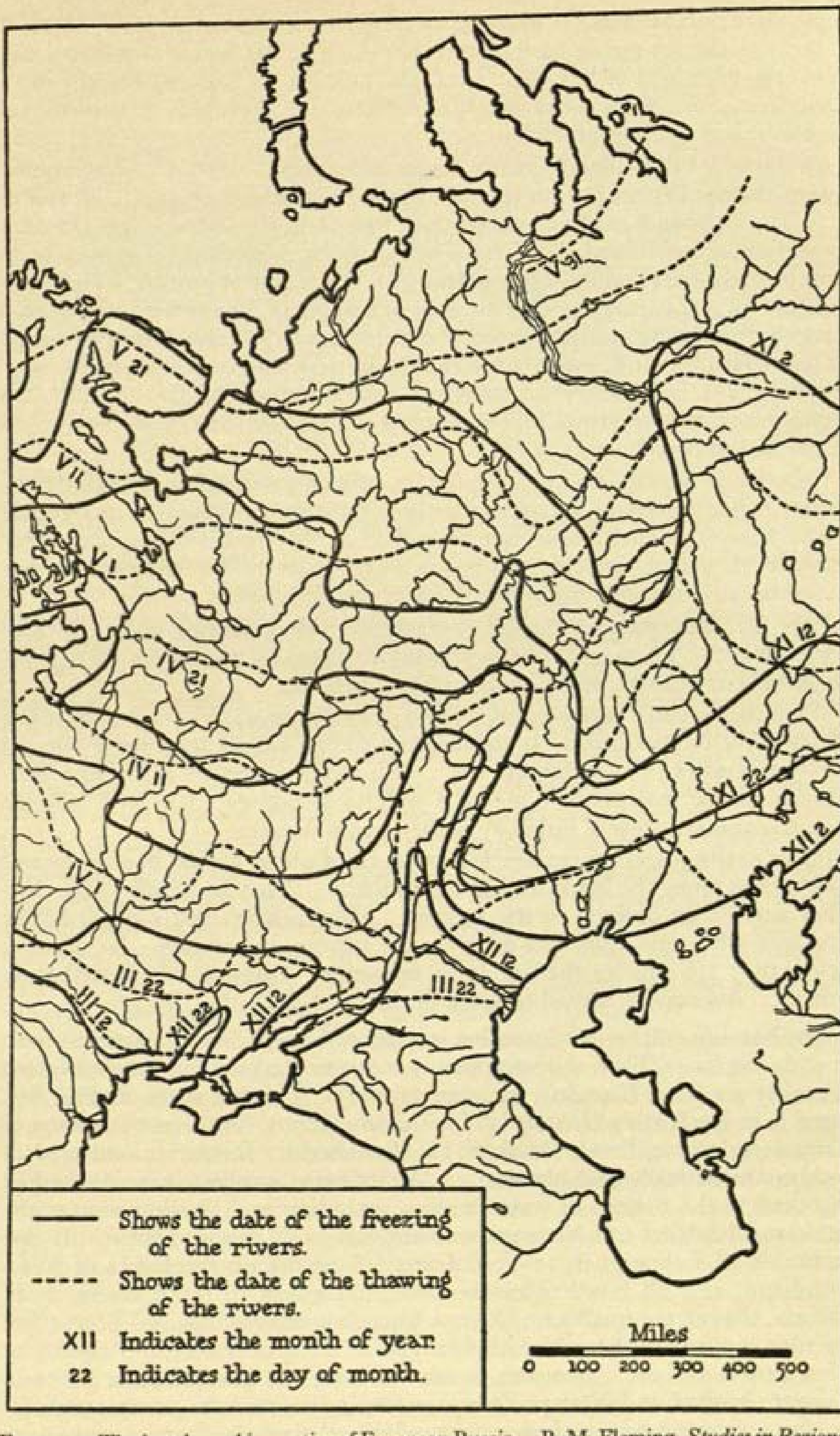


FIG. 45.—The isotaks and isopectics of European Russia. R. M. Fleming, *Studies in Regional Consciousness and Environment*, 1930, p. 5, map. 1.

along the principal axis,[330] because there is free space for crystallisation,[331] or because the ice grows under pressure.[332] The vertical arrangement explains the high load of lake-ice,[333] viz. 260 lb/sq. in., as against the 185 lb in glacier-ice whose plates are variously orientated. B. Weinberg[334] determined the coefficient of internal friction.

Cholnoky[335] described the early stages of lake-ice. In quiet, supercooled waters, the needles are horizontal and disposed in feathery aggregates at angles of 30°, 60°, 90° and 120° with each other, though in the initial stages they are not definitely orientated.[336] Those below the surface thicken and unite to form pancake-ice (Forel's *glaçon-gâteau*[337]). The rate of growth, which controls the air-content, is quicker than in sea-water in which the sinking cold solution, formed by freezing, forces the warmer water to rise. It is governed by temperature, wind, snow-cover and cloudiness, and is at first rapid, later less rapid—it is inversely proportional to the square of the time.[338] In Alpine lakes, the ice grows to a maximum thickness of 80 cm, in Norway to 70 cm and in Sweden to 1 m.[339]

Lake-ice is warmer than glacier-ice by an amount which increases downwards. Cold waves do not lower the temperature so much as initiate freezing at the base, the latent heat warming the ice.[340] Stefani showed mathematically that the cover can thicken for a time after the low temperature has ceased because the cold wave continues to pass through it.

Lake-ice undergoes no change of structure unless melting and refreezing supervene.[341] These give rise to transitions between water-ice and glacier-ice and even, as some assert,[342] to true granular ice.

Warming by sun's rays produces Tyndall's melt-figures[343] (see p. 58) in the planes of the impure layers. It disintegrates the ice into its separate prisms,[344] the "ice-candles" or "prismatic ice". The falling asunder is owing to impurities in the interprism films—the difference between sea- and lake-ice from the point of view of impurity is only one of degree.

Plates in river- and stream-ice[345] form at the surface, at first in shallow and quiet places along the banks and about boulders. They lie parallel with each other and partly oblique to the surface. They thicken and build a lattice structure with new ones that grow in the intervening spaces. In running water, they are smaller than in lakes or ponds. Streambanks retard their growth. Anchor-ice furnishes some of the crystals.

Anchor-ice. The ice-formation is quite different when the streams have an eddy motion. Then the ice grows as swimming crystals and at the bottom as anchor-ice (Ger. *Grundeis*; Fr. *glace de fond*). This ice, since R. Plot first noted it in his *Natural History of Oxfordshire* (1677), has been the theme of many descriptions[346]: O. Devik[347] has discussed its formation and growth both mathematically and physically. As its name implies, it is attached or anchored to the bottom of water-masses, including such shallow seas as the Baltic and Kattegat and Norwegian coastal waters[348] down to 60 m. It also occurs[349] off Labrador in 10–15 fathoms (18–27 m), on the Banks of Newfoundland, and off South Victoria Land. It grows in rapid rivers, as in Canada, though the frazil ice of French Canada is not ground-ice[350] but a fine spicular or disc-shaped surface-ice formed in streams flowing too rapidly to form a surface-sheet. It occurs in north Russia and north Siberia (Irtysch, Taimyr, Anabar) and Novaya Zemlya, where swift currents prevent surface-ice from growing but anchor-ice may choke the whole channel and cause

flooding if the cold is prolonged. It is also important in connexion with power-plants (see below).

Anchor-ice is a spongy mass (in appearance not unlike frog spawn) of intermingling, spicular crystals and partly bedded thin plates or rounded discs. The ice-particles occur as colloids in tiny, rounded plates[351] which adhere by regelation. They grow up to 1 m thick in arborescent forms on days with strong cooling but build more massive and compact layers on days of moderate cooling and slower production. Such observations as have been made prove that a thickness of 0·75–1 m or even 1·5 m may grow during one night.[352] Growth from the bottom increases until the lifting power due to the lighter specific gravity, aided by the penetrating sun's rays at daylight which warms the dark rocks, overcomes the cohesion of the ice to the bottom so that boulders, etc., are raised to the surface: lost anchors appear in this way. The ice bursts up with seal-nets and ropes or with a load of seaweed, frozen mud, sand, shingle or slabs of rock, especially in early mornings following clear, cold nights. Boulders are upheaved if they rest on other boulders or if their base of support is narrow.

M. Arago[353] thought anchor-ice only grew in slowly moving waters at the bottom of a stream on pointed boulders or pieces of wood, mouths of fish, etc., which acted like ice crystals in an oversaturated solution, a view later developed by W. Röntgen[354] and G. Tammann.[355] Growth is only possible in running waters since only these can be cooled to the bottom, partly by conduction and radiation,[356] principally by turbulence[357] which in a few minutes transmits the cold to the bottom and causes a no less rapid carrying off of heat from the surface. The vital part played by turbulence is proved by experiment.[358] Undercooling is apparently a necessary condition[359] though some writers deny it[360]: the amount in Russian rivers was 0·001°C, exceptionally 0·1°C,[361] though in laboratory experiments, in which seeding by ice-crystals is carefully avoided, supercooling may be 5–10°C.[362] The connexion with the specific heat of bodies at the floor was shown by Keever and proved experimentally.[363] In hydraulic power plants anchor-ice is prevented from forming by heating the grids in the front of the inlets to the turbines and by streamlining the inlet passages to reduce turbulence.[364]

Anchor-ice is not confined to fresh water. It is found in seas (see above) in which the overcooling depends upon the salinity and possibly upon currents of different salinity[365] or upon emergent springs.[366]

G. Lussac[367] attributed the formation to small ice-needles descending to the bottom where cooling caused them to adhere. This view, though not unsupported,[368] is improbable[369] since surface ice-needles are not observed in such situations nor are they heavy enough to make the descent. Some writers are of the opinion that the ice may crystallise anywhere in the water.[370]

Anchor-ice is unknown beneath bridges, overhanging rocks, trees, surface-ice or other obstacles to radiation[371], or on sunny days or cloudy nights. In the Antarctic, however, it is independent of the state of cloudiness, the colour and constitution of the object on which it grows, and the freedom of the water from ice, the overcooling being brought about by other means.[372]

The conclusion that anchor-ice grows at any depth, provided the vertical temperature distribution permits cold wedges to form between two warmer layers,[373] finds a basis in O. Petursson's discoveries in the Kattegat.[374] Observations on Russian rivers[375] suggest, however, that intermingling maintains an energetic exchange of heat between the cooled surface and the

bottom, the latter without appreciable radiation, thus affording a means for continuous supercooling and crystallisation.

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PART II

GLACIAL GEOLOGY

(*A*) *GLACIAL EROSION*

CHAPTER IX

GENERAL PROBLEM

About one-tenth (= 10·7%)[1] of the total land-surface of the globe is at present glacierised and about one-third was glacierised during the Glacial Period. These figures give a measure of the importance of the glacial processes and their morphology in one of the principle fields where stratigraphy and physical geology meet.

History of views. Attrition by moving ice, connoted by "exaration",[2] "glacerosion"[3] or other words[4] not so good as the general word "erosion"[5] ("corrosion" has been employed to denote lowering by glacier or river[6]), has been the theme of prolonged debate. Its recognition followed long after that accorded to rivers. Failure to appreciate the glaciers' physiographic role was because their extent was more limited and their action less obtrusive, though often erratics are impressive and drifts immense. Effective glacial erosion was only tardily accepted: it is still stoutly contested.

In the early days of geology, when under the belief in the Deluge erosion of any kind was not well assured, the glacier's share in the origin of valleys was naturally not particularly considered. Ice-marks only drew the attention of geologists and Arctic explorers as signs of previous glaciation. Forbes,[7] for example, confined himself to pure description. Yet F. Le Blanc[8] recognised the glacial relief or glacial *Formenschatz*, as it later came to be called (see p. 227), and distinguished regions which are or have been glacierised, and J. A. De Luc[9] and N. Desmarest,[10] observing glacier-debris and the rounded blocks in Swiss moraines, almost simultaneously (in 1779) concluded that ice eroded. But while De Luc thought ice was essentially conservative, J. Esmark[11] (1827) ascribed to it the Norwegian fjords and Agassiz[12] (1837), who thought it rounded the angles and planed the surfaces, judged its power to be considerable. Thus early appeared the cleavage of opinion that still persists.

The following decades, devoted mainly to ice-structures, causes of ice-flow and the extent and direction of movement of the Pleistocene ice, had only an occasional advocate of ice-erosion[13]; Esmark, as we have just seen, invoked this action for fjords, as did Dana,[14] while J. Y. Hind[15] postulated it for the Great Lakes. The main controversy opened with Ramsay's papers[16] (1862, 1863) which maintained that ice hollowed out rock-basins. Ramsay's view, which could only have been promulgated at this time when the Pleistocene glaciation of mountain regions had won general acceptance (see ch. XXX), found much favour (see p. 265). Extreme protagonists extended it to cirques[17] and U-valleys (see p. 319). The controversy, after a lapse of a few decades, was renewed about the close of the century when morphological studies led A. Penck[18] in the Alps and E. Richter[19] in Norway to postulate much erosion. They caused the tide of opinion, which has at times supported the doctrine of profound glacial erosion and at others discredited it, to set in

strongly in favour of the theory. When revived, this took the severer form of the U-valley, as advocated in North America[20] by W. M. Davis and G. K. Gilbert and in Europe by Penck[21] who raised many new problems and awakened fresh interest. These glacialists demanded a degree of erosion which far surpassed the claims generally made in Ramsay's day: the focus of attention shifted from rock-basins, the subject for example of J. Geikie's various essays on ice-erosion (see p. 265). Yet, there is to-day a distinct reaction from the views that prevailed at the opening of the century. Once more the pendulum of opinion has swung from the theory of excessive glacial erosion to a position perhaps midway between the extremes: it cannot yet be said to have come to rest.

Two schools. There have been two schools of thought, diametrically opposed, since glaciers were first recognised as an erosive force: the one credits glaciers with virtually no such powers, the other raises them to be the severest denuding agent known to geology. The conflict, it has been said, is between glacier experts and physiographers. While few geologists[22] who have laboured in the glaciated territories of North America or north Europe have assigned to ice a conservative role, almost all glacier experts have believed that ice was essentially incompetent and merely put the finishing touches to a pre-existing topography. They have included Alpine glacialists,[23] e.g. L. Agassiz, E. Desor, A. Escher, B. Studer, A. Heim, E. Richter, L. Rütimeyer, F. A. Forel, J. Sölch and R. v. Klebelsberg, and those familiar with ice in the Alps and elsewhere,[24] e.g. C. Martins, T. G. Bonney, E. J. Garwood, J. W. Gregory, M. Conway, D. W. Freshfield, S. Hedin and E. Whymper. Early French geologists, as É. de Beaumont, E. J. d'Archaic, E. P. de Verneuil, A. Daubrée, E. Collomb and C. Grad, almost unanimously shared this opinion.[25] J. Ruskin[26] expressed it in its extreme form: ice acted as a sponge to remove what water chiselled; it could no more scoop out basins than the custard could deepen the pie-dish. Interglacial epochs, it is said,[27] were periods of valley excavation, glacial epochs times when land forms were conserved. Glaciation means relative cessation of valley formation[28] (*Gletscherperioden sind Ruheperioden*) since the ordinary natural processes are arrested or in abeyance, the rocks being shielded from frost, from appreciable changes of temperature, from alternate wetting and drying, from avalanches and from chemical and plant action. "Ice is nature's substitute for sand-paper."[29]

Antarctic ice is no exception[30]; it has a low velocity (see p. 104), few streams and little frost action or debris.

"Glacial protectionists" discriminate between the transporting and erosive powers of ice,[31] conceding great efficiency to the one (blocks of 8000 cu. m have been seen in transport) and denying it to the other.

The other school of disputants has credited to ice the major physical features. Ice scoured out such sea-passages and inlets as the North Channel,[32] Menai Strait,[33] Bering Sea and Strait,[34] Cabot Strait,[35] Bay of Fundy,[36] Norwegian Channel[37] and submarine canyons.[38] Entire valley systems, like the Irish glens,[39] the Norwegian fjords[40] and the valleys radiating from the Fens,[41] are but gigantic grooves which the polar ice engraved on a magnificent scale. Ice completely obliterated the mature preglacial relief and drainage plan of the western Alps,[42] planed off the summits and crests of the Vosges,[43] and wore down the great plains of Europe and Asia,[44] including the plateaux of south-west China. It evened the submerged ridges between Scotland,

Faeroes, Iceland, Greenland and Baffin Land,[45] broke the continuity of the continental shelf of Norway,[46] and created the shelf's low level off all glaciated coasts.[47]

In the theory's most fantastic form, the universal ice scooped out the Red Sea, Arabian Sea, Persian Gulf and even the ocean basins.[48] Only less modest was the claim that the prodigious erosion caused the successive ice-sheets to decrease progressively[49] and finally to disappear.[50] By lowering the surface of the country they warmed the air, curtailed the snowfall and starved themselves: they dug their own graves.

These extravagances require no refutation. But between the "glacial ornamentation" of the glacier experts and the "glacial architecture" of workers in glaciated lands, e.g. E. Brückner, M. H. Close, J. Geikie, A. Helland, J. B. Jukes, A. Penck and A. C. Ramsay in Europe, T. C. Chamberlin, R. A. Daly, W. M. Davis, H. Gannett, G. K. Gilbert, W. E. Logan, J. S. Newberry, R. S. Tarr and C. Whittlesey in North America, and J. v. Haast and J. Hector in New Zealand, there are all manner of gradations. No wonder this question has occasioned a wider divergence of opinion and consequently a more voluminous literature than any other branch of glacial geology: it still remains a vexed problem.

Nevertheless, unanimity now obtains on two points: first, that mountains, plateaux, escarpments, major valleys and river systems are, with other leading features, preglacial in date and origin; secondly, that the ice stripped off the preglacial soil, weathered material and mantle rock, notably on the uplands, and rubbed down minor inequalities, in the process striating, grooving and polishing the rocks, rounding the hills, and broadening the valleys. The question at issue is narrowed down to the smaller one of magnitude and intensity, namely, in what ways were such intimately related features as rock-basins, cirques, U-valleys and fjords eroded? (see chs. XII to XV).

Difficulties. Formidable difficulties stand in the way of a solution; for the problem contains many conflicting and uncertain elements. Apart from a few inconclusive figures (see p. 222), no quantitative results are available. Rock-surfaces, glaciated under the thick and freely moving ice, have been later subjected to the modifying action of the thin and dwindling snouts.[51] They are often masked by screes, glacial deposits or standing water. The sole of modern glaciers, the seat of erosion, is only open to inspection at the bottom of bores, e.g. in the Hintereisferner,[52] in shallow crevasses, e.g. near the Jungfraujoch,[53] in ice-free windows, e.g. in the Rhône ice-cataract,[54] in tunnels driven through some Italian glaciers,[55] in moulins (Carol, 1944), or at the margin—it was said in 1907 that the total distance man had penetrated beneath the world's glaciers did not then exceed 500 m.[56] In the Antarctic, the larger of the two modern ice-sheets, the edge itself is mostly inaccessible.

Contradictions occur in everything pertaining to ice and its work. Rigidity is opposed to plasticity; and incoherent accumulations are severely folded and eroded or are overridden without disturbance (see p. 218). Although these manifestations are certainly reconcilable—some indeed are associated with running water or other geological agents—they inevitably throw the argument back upon our conception of the physical conditions at a glacier's base and upon our interpretation of glacial topography. Hence, different aspects and arguments have borne the emphasis during the various phases of the controversy. The subjectivity of much of the evidence has caused

assertion to stand against assertion and has split opinion about the rate, amount and method of glacial erosion.

The loose meaning attached to the very words ice-erosion has introduced unnecessary confusion. Is a glacier's competency to be measured by the absolute amount of the erosion or by its ratio to what other subaerial agencies like rains and streams might have accomplished in the same area over the same period?

An appeal to modern glaciers, unfortunately, is of little avail, since at their margins loads are excessive (see below), thickness and motion are at a minimum, and the soles are at melting point and commonly separated from the ground beneath by hollows, caves and channels. Both the Greenland and the Antarctic ice-sheets are thought to be protective[57] (see p. 212). Further, the glaciers are generally in a recessive phase (see p. 146) and repose in valleys which are broader than their present dimensions require: "lateral moats" many hundred metres wide separate the Ferrar Glacier from its rocky walls.[58] Manifestly, evidence of glacial erosion is no more to be sought here than is proof of strong fluviatile erosion to be looked for on desert plains where rivers dry up. There is, indeed, practically no glacial erosion in these positions[59]—an advance left intact the chisel marks in a marble quarry at the foot of the Lower Grindelwald Glacier[60] (cf. p. 222).

Methods of erosion. While some early writers like Charpentier and Agassiz attributed any erosion there might be to the ice itself, others directed attention to the weight of the embedded boulders.[61] That clean ice *per se* can remove only loosened waste and trivial amounts of live rock has long been admitted[62]; its relative softness is adequate justification if we remember that a glacier's base, except near the margin, is at or only a little below 0°C (see p. 107) and that the hardness of ice in Moh's scale varies as follows[63]: a few degrees below 0°C, 1–2; −15°C, 2–3; −40°C, *c.* 4; −50°C, *c.* 6.

The higher and colder layers do little[64] since they are fairly clean or hold their debris insecurely. Although the latter makes ice an efficient agent, the erosion is raised up to a certain point only beyond which it is diminished by retarding the flow. Discussing the effect of this ratio, I. C. Russell[65] showed that abrasion is severest if the bottom layers are lightly charged and that one of the factors diminishing it towards the snout is the excessive load of basal and englacial debris which, as elsewhere, may lead to immobility (see p. 230).

The mechanism, though not fully understood, involves wear by subglacial streams, especially near the margin (see ch. X); abrasion, including striation and polishing (see ch. XI); plucking along structural planes (see p. 249); overcoming the cohesion of the cementing matrix in rocks by ice-pressure (see p. 251); changing pressures at the sole (see p. 301; thrusting of boulder-clay into joints and other planes (see p. 364); and infiltration of muds and subglacial waters, including those derived by earth's heat, under the great ice-pressure into the rocks which they attack chemically and mechanically.[66]

Several geologists,[67] including Ramsay, stressed the weight of the ice. But mere weight is unavailing since it is less than the crushing strength of rocks[68] —ice, 300 m thick, exerts a basal pressure of *c.* 300 bars, i.e. only about one-hundredth of the pressures commonly used in P. W. Bridgman's laboratory. Hence, rock-crushing is either non-existent or quite trivial,[69] though it influences plucking when combined with flow. The latter is the vital factor as Tyndall recognised; its power may increase with the third power of the basal velocity[70] or more probably, as experiments with other

solids show, approximately with the first power.[71] Little depth and high velocity may accomplish as much as great depth and small velocity.[72] Since plucking exceeds abrasion (see p. 250) and is encouraged by small thickness and great flow-pressure (see p. 250), erosion may indeed be severest if the ice is not very thick. A critical mass and velocity are seemingly essential,[73] erosion being little up to a certain velocity but effective above this. Antarctic glaciers, flowing annually 30 ft (*c.* 9 m), may be below this velocity,[74] unless they move over soft or jointed rocks. While static ice can only be protective, glaciers which have a block-movement (see p. 118) may erode severely[75]: fjord-basins may have been hollowed out in this way.[76]

Regional distributions. In examining the problem of erosion, certain regional differences must be borne in mind. The complementary processes of destruction and construction had their maxima in different areas; the first acted mainly in central, mountainous regions where flow was rapid and the disintegrated products were quickly removed, the second over the low periphery where the ice was thin and where sluggishness or stagnation weakened its power.[77] This crucial difference was only fully manifest if the ice could spread freely over low lands, since relief was largely in control over dissected uplands.[78] These distributions which, be it remarked, are not those of nourishment and dissipation of the ice itself, are particularly well seen in north-west Europe where the frontier between "exaration" and accumulation bisects the Baltic.[79] The erosive centre comprised Scandinavia and the region about the Gulf of Bothnia, as the Swedish and Finnish coastal directions and outlines prove, together with the area north of a line running from Courland to the north-east corner of Lake Peipus.[80] Drifts of any thickness are comparatively rare in north Norway and cloak scarcely one-tenth of the country.[81] They occur however at Andö[82] and extensively in Lister and Jaeren in the south-west where they give rise to a type of scenery which is reminiscent of the North German Plain rather than of Scandinavia. Finnish lakes[83] are chiefly rock-basins, though drift determines most of the lake-contours. The average thickness of the drift in Norrland is 4–7 m.[84] In the Canadian Shield and Barren Ground of North America,[85] roches moutonnées, rock-basins and rapids abound and rivers appear unfinished.

The peripheral zone of aggradation (A. G. Werner's *aufgeschwemmtes Land*) comprises the great north European plain, Denmark, south-east Sweden, Lithuania, Latvia, north-west Russia and the country east and south of the White Sea. The production of thick drift over much of this belt was furthered by the softness of the solid formations, e.g. Old Red Sandstone in the east Baltic, Tertiary strata in north Germany, and Tertiary and Cretaceous rocks in Denmark, as well as by the prolonged halts and marginal ablation that accompanied the recession. In this zone, striae are extremely rare, since exposures are few and the rocks unsuitable, e.g. Mesozoic and Tertiary clays in north Germany and East Anglia and Mesozoic strata in South Wales. They are very rare, for instance, in Denmark (*cf.* map and list[86]), over the North German Plain (cf. list and map[87]), in Poland, Lithuania and Russia,[88] and over the English Midlands.[89] The control was partly lithological and partly glacial.

The same is true of the corresponding belt in North America, namely, the plains of west Canada and the land south of the Great Lakes, where soft Palaeozoic and Mesozoic rocks prevail—in the Lake Simcoe district of Ontario, the boundary between the ice-erosion area and the drift area

coincides roughly with the Pre-Cambrian–Lower Palaeozoic junction because of the difference in lithology, the soft Ordovician rocks being readily eroded.[90] This is seen, for example, in the short list of striae compiled for New York State.[91] Soft sediments yielded abundant material for the stout moraines and thick drift sheets of the Middle West. East of the Great Lakes the drift becomes thinner and less continuous with the occurrence of more resistant types of rock.

Thick drift swathes the continental shelf off the glacierised lands of to-day and the glaciated lands of the Glacial period, as off north-west Europe, Greenland, Patagonia and eastern North America.[92] The shelf is indeed largely encumbered with terrigenous waste, much of it glacial drift,[93] for example, off west Norway[94] where the drift consists of angular material containing much felspar and builds great shore banks and probably constitutes the flat, submarine floors outside the Skjaergaard and extends even down the continental slope—submerged moraines have been recognised.[95] A like opinion may be held for the Barents Sea,[96] whose floor is strewn with rock-fragments; for the Irish Sea where glacial erratics abound[97]; and for the Dogger Bank and shallow parts of the North Sea[98] which have extremely few solid outcrops, have yielded a varied assortment of erratics and are swathed in drift which probably has largely filled in the preglacial and glacial river-channels (see fig. 51, p. 256). In marked contrast, the floor of the English Channel which lay beyond the reach of the ice is mainly solid rock. The even floor of Hudson Bay, which owes its existence to ice-action,[99] may be a consequence of glacial deposition.[100] Nevertheless, the basins, troughs and banks of the shelf off glaciated lands have been interpreted as products of ice-erosion.[101]

This distinction between a central area of erosion and a peripheral one of deposition is only broadly true. There are indeed four regions: a central one of non-erosion, a subcentral belt of maximum erosion and a submarginal belt of maximum deposition grading into a fourth or marginal zone of little deposition.

That the centre, situated under the deepest ice, did not have the maximum erosion was due to the purity and feeble flow of the ice (see p. 124)—the ice, it has been said,[102] rested on pressure-water or sludge ice. It has almost no signs of ice-erosion; erratics are local and angular, striae are few and faint, and valleys, as on the Norwegian ice-shed, are but little modified.[103] Near the Keewatin centre, as A. P. Low was the first to notice, rock-fragments are angular and if rounded are the cores of rotted granite and gneiss.[104] Gneiss in Labrador[105] is not worn down to the level of the preglacial decay in its intersecting dykes; its hills are sharp and angular; and the Pre-Cambrian surface, exposed to the Pleistocene ice, generally resembles that which the Palaeozoic cover has protected. Elsewhere in the Canadian Shield, an intricate pre-Ordovician drainage pattern, closely adjusted to weak belts in Pre-Cambrian rocks, remains unaltered by glaciation save for some shallow rock-basins.[106] The valleys and central tablelands in British Columbia, which were under the Cordilleran ice, have likewise suffered little[107]: their very glaciation in certain parts was questioned for a time.[108]

This is also true in the equivalent European areas. It was noticed, for instance, as early as 1845[109] that the *Friktionsphaenomenet* was less intense on the central Scandinavian plateaux than along the fjords and margins of the peninsula. Preglacial gravels and rocks, disintegrated down to as much as 10 m or even 52 m, remain untouched on parts of the ice-divide in Lapland[110];

deposits of impure kaolin formed by secular decay occur in east Finland[111]; weathered gossan from the preglacial zone of decomposed ores persists as boulders in north Carelia[112]; and lakes are almost absent.[113]

An earlier relief is locally preserved in the central[114] and preglacial breccias in the eastern Alps.[115] Remnants of a Pliocene weathered crust have persisted in the heart of the Black Forest's glaciation[116] and a preglacial slip has survived in Mull.[117]

Erosion was small where the local ice was unable to escape,[118] as in the Howgill Fells of England and in places in the Southern Uplands of Scotland (the overlapping spurs are intact), and was at a minimum in the ice-logged, diamond-shaped areas of flow (see p. 715). Such critical ice-relationships helped to preserve preglacially weathered Moine gneiss north of Blair Athol in the Scottish Highlands,[119] weathered rock in eastern Scotland and north-east England (see p. 224), Lough Neagh clays in northern Ireland, and deeply weathered rock at the boundary of the Keewatin and Patrician ice-sheets in south Manitoba.[120]

Erosion was most severe in the subcentral zone of high floor-gradients, of deepest and most quickly moving ice and (save for the central area itself) of the longest glaciation. It was probably inframarginal,[121] say, a few hundred miles or kilometres back from the edge, as the relative roles of abrasion and plucking and the latter's severity under thin ice with an effective flow (see p. 215) seem to warrant—the maximum flow was under and downstream from the névé-line.[122] Yet deposition was not completely lacking. For instance, lakes in Lapland are mainly stau-lakes[123]; and drift, exceptionally 12 m thick but averaging a few metres, shrouds four-fifths of Finland[124] (see below) and is distributed over one-quarter of the Alps.[125] Yet the drift, as on the Laurentian plains, usually occurs in more or less sheltered valleys and in the lee of rocky eminences.

Planation was least at the periphery. Here thick drift frequently hides layers of subaerial wash and local debris[126] and interglacial accumulations and earlier drifts whose forms have been preserved. The Nuneaton Cambrian quartzites have jagged edges.[127] The topographic features within the drift, e.g. in Ontario,[128] resemble those of stream erosion in unglaciated regions and, in marked contrast with the central area, are not definitely orientated with the ice-flow. Sand, gravel and clay mantle more or less continuously the extensive plains of this zone: rock is exposed only in occasional inliers or in the walls and floors of glacial epigenetic gorges.

The zone of maximum accumulation in north-west Europe coincides with the Baltic Ridge of north Germany and Poland, as elucidated by isopachytes[129] (see p. 360), and extends into the plain north of the watershed of south Poland[130] (up to 250 m thick), into Estonia and Latvia[131] and into East Anglia.[132] West of the North Atlantic, it runs south of the Great Lakes under the prairies of the Upper Mississippi basin and the central plains of Quebec. Greenland, which possesses few glacial accumulations, has no equivalent (see p. 216).

The drift thins fairly rapidly towards its boundary where it becomes patchy and is represented by an occasional erratic. Thus in Europe it thins through Belgium towards the coast[133] and from East Prussia, where it is 200 m thick, to 100 m near Lodz and 50 m near Kalicz,[134] becoming negligible south of the main Polish watershed.

At the margin, the ice being thin and sluggish, the difference between

glaciated and unglaciated country almost vanishes, as revealed for instance in the rock-contours along this line in New York State.[135] This accords with the superimposed and undisturbed drift sheets and loess horizons in the Mississippi region, and with observations in the outer region of the drifts in Pennsylvania where the ice had little or no movement[136] and at the glacial boundary on mountain slopes where, as in the Alps (see p. 40), the erosion did not surpass small and assignable limits.[137]

This fourfold distribution was modified as the ice-centres were displaced (see p. 670) or the ice-sheets waxed and waned. The zones of preponderant erosion and accumulation migrated, advancing and receding with the ice. Every locality, except at the outermost limit, was invaded twice, with interglacial movements, several times, and placed in the marginal zone. Most of the erosion on the Canadian Shield occurred while the submarginal parts of the ice were expanding across the region during each glacial epoch and most of the drift in the belt of maximum erosion, as in Finland, originated during the recession.

Methods of approach. Although the problem of ice-erosion has been studied continuously and from many angles, relatively few facts have been tendered in proof or disproof. The literature is largely a repetition of unfounded assertions or of indirect arguments of doubtful validity, conclusions varying with the predilections and field experience of the disputant.

The methods of approach are four in number[138]: (1) by examining the visible processes at work in connexion with modern glaciers, as when these advance on to loose material, or by comparing glacier-muds with those of neighbouring streams and rivers—this method is not very fruitful and has often led to adverse and partially erroneous judgments; (2) by contrasting glaciated with unglaciated relief and ice-worn with water-worn topography—this physiographic and most important method, which is complicated by the activity of associated agents, has usually exaggerated a belief in the erosive severity of ice; (3) by deducing the appropriate consequences of both the affirmative and negative propositions and confronting them with the facts in glaciated regions[139]; (4) by applying theoretical principles of the mechanism of glaciers—this useful auxiliary has helped little in the past.

Action on incoherent material. Adherents of the two schools, mentioned earlier, have often appealed to the glacier's action on incoherent materials along its oscillating edge. Charpentier[140] noticed that glaciers sometimes ploughed up moraines, an observation others often repeated in subsequent years.[141] The ice ploughs up fields, uproots and fells forests, and pushes up the sea-floor and its shells into terminal moraines (see p. 631). It flutes and corrugates sands and gravels parallel with its flow and to a depth of a few metres[142]; fashions forms resembling roches moutonnées[143] in outwash or interglacial gravels; and damages constructions erected by man.[144]

Arguments founded upon these occurrences are offset by others of contrary nature. Charpentier[145] himself recorded how the Glacier du Tour overrode gravels for a distance of 80 m which, with their tufts of Alpine plants, were seen to be undisturbed when the ice receded after five years. Similar observations had already been mentioned in an old Grindelwald chronicle of 1588 and by L. v. Buch[146] in Norway. They have been often and more recently reported from various glacier regions,[147] including Alaska: the retreating Muir Glacier uncovered a cedar forest, erect and in place.[148]

Granite boulders, wrapped in sands of their own decay, have been left untouched.[149] Such overriding, noticed too in connexion with fluvioglacial gravels of the early Pleistocene,[150] is sometimes accompanied by the snout's elevation into a raised lip or an anticline.[151] It has been reproduced experimentally[152] as in W. J. Sollas's "pitch glaciers".[153] Glaciers may likewise move bodily over inequalities with only partial adjustment or may be turned aside or be split by their own moraines.[154]

A closely related phenomenon is the fluted sole of the ice which may extend 9 or even 21 m in the lee of an obstructing boulder or rock-projection.[155] Fans of debris in the lee of boulders in the drifts below the Upper Grindelwald Glacier may be analogous.[156]

Contradictory too are the absence of any trace of ice-erosion on solid rock recently uncovered by the ice[157] and the striations or minor plucking sometimes seen in these positions (see p. 222).

This striking contradiction in glacial behaviour, an expression of the brittle and plastic properties of ice, may be explained by the irregularity of the floor[158] but more probably by the glacier's phase[159] which reacts by controlling the rate of flow and the relative velocities of the upper and lower layers. Ice in the lee of boulders may also be fluted because these, warmed by the sun's rays, melt their way into the sole.[160]

Overriding is associated with thin margins. It no more disproves ice-erosion than does the fact that water, the erosive power of which is unquestioned, works in the same way. In dry seasons, rivers flow gently over their sandy beds without disturbing them; erosion is at a standstill. To-day, the world's glaciers, even in polar latitudes, are in a drought stage compared with the Pleistocene[161]: Himalayan glaciers are often riding on the top of a rampart of ground-moraine.[162] Consequently to judge them by phenomena at their snouts, precisely where erosion must be weakest, is as illogical as it is to appraise the performance of rivers at the moment when they lose themselves in a desert. "I have long congratulated myself that I was not trying to convert a sceptic by showing him what a glacier in full activity can do", said Sir C. Lyell.[163]

Glacier-muds. Glacier-streams, as Scheuchzer[164] remarked in 1723, are thickly loaded with fine mud which renders them turbid and milky white (hence the Icelandic word *Hvitá* ("White") in river names and the names *Weisssee*, *Lac Blanc* and *Lej* of Alpine *Hochgebirgsseen*). The "brown zones" in front of tidal glaciers in Alaska, Greenland and Spitsbergen, which have great zoological significance, are connected with the upwelling of muddy waters.[165] The particles are so minute and finely powdered that a sample of the water takes several days to clear—they remain in suspension even after passing through the long settling basins of lakes Pukaki and Tekapo in New Zealand.[166] The streams are only limpid if, for instance, the ice is practically stationary, e.g. in places in Greenland[167] or the Antarctic (see p. 418), or if they filter through fine materials[168] like sand and gravel or Icelandic tuff.

The pipe-clay colour, early imputed to flow over white limestone or marble,[169] Forbes correctly attributed to the action of glaciers upon their beds.[170] The impalpable powder, some of the order of 0·0005 mm,[171] is the ultimate product of their scour. By determining its amount we may get some conception of the scale of their erosion.

The colour and opacity of glacier-streams increase with the volume of water to daily and summer maxima, though the dissolved matter varies less than

the suspended matter.[172] Estimates on the Lower Aar Glacier[173] gave a daily carriage of 280 tons of sediment. Comparable figures have been gathered from the Rhône and Mont Blanc glaciers,[174] from Norway[175] (Jostedalsbrae transported 1968 tons in one day), the Sareks region of Lapland,[176] Greenland,[177] Iceland[178] (Vatnajökull streams remove annually 15 million tons), and from Alaska, the Muir Glacier yearly producing 15 million cu. m.[179]

Since satisfactory figures must be founded upon measurements taken daily over several years, the above estimates which do not comply with these conditions serve to demonstrate only that glacier-streams are transporting enormous quantities of sediment. But calculations based on these or similar crude figures have been made of the rate at which glaciers erode their beds. Omitting the coarser detritus, the annual lowering by the Lower Aar Glacier was found to be 0·6 mm.[180] Minimum values[181] gave the Reuss 0·24 mm, the Rhône 0·29 mm and the Kander area 0·43 mm; the lowering in Paznaun was 0·113 mm.[182] Corresponding figures[183] for Jotunheim were 0·54 mm (or 2·5–3 mm for part of it), for Hardangerjökull 0·69 mm and for Jostedalsbrae 0·079 mm. The mean annual lowering by the lobe of Svartisen, calculated from the rate of infilling of the lake at its edge, was 11 mm.[184] The Sareks region yielded 0·5 mm,[185] the Karsa Glacier *c.* 1·5 mm,[186] the Vatnajökull 0·647 mm,[187] the Hoffellsjökull 2·8 mm[188] and the Muir Glacier 0·02 mm.[189] The general rate of lowering has been estimated at 0·5 mm/annum or that of running water.[190]

Hess,[191] employing a different method, namely, by measuring the debris that melted out of the Oetztal Glacier along its median moraine and allowing for the share superglacially contributed in the firn according to Finsterwalder's theory (see p. 119), computed the annual lowering at 0·027 mm. He also obtained 0·005 mm for the entire basin of the Hintereisferner and 5 mm along the axial line.

These and like calculations err in two opposing directions. They undervalue the erosion by ignoring the more heavily laden deeper waters and the rolling "bottom" load of gravel and shingle whose importance valley trains and outwash fans indicate. They go astray in the contrary sense—and possibly appreciably—by including mud the ice did not contribute[192]; some of the mud was gathered from surface-moraines and by meteoric forces, by the slipping of moraines and glacier-tables, by atmospheric waste from rock-walls and cliffs, particularly in the firn, and by the mutual attrition of stones in crevasses. A higher fraction, possibly more than half, may have come from subglacial streams working upon subglacial rocks and fragments in suspension.[193]

Yet glacier-muds unquestionably testify to wear by ice; turbid glacier-streams contrast with the limpid waters from adjacent ice-free valleys,[194] notably in winter (this is disputed[195]); the glacier-milk is related to the intensity of glaciation[196]; the grains of sand in glacier-streams are angular[197]; the mineral particles are fine and have been ground mechanically from the rocks, and in waters containing CO_2 have been derived by chemical processes involving hydrological changes.[198]

Nevertheless, glacier-milk is but an insecure foundation upon which to base any quantitative conclusions,[199] especially if the data have not been collected by continuous readings taken at all seasons and from the whole basin.

Subglacial erosion is also obviously demanded for the glaciated superglacial material that occurs where, as in east Greenland,[200] the rocks are

completely buried; for the gneissic erratics on the volcanic rocks of Gaussberg[201]; and for the inner moraines which are found where superglacial moraines are wanting,[202] as in the small Pyrenean glaciers or the east Alpine Schwarzensteinkees. Equally conclusive are the erratics from the Silurian Limestone of Gotland and Ösel and the chalk and basalt of Scania (see p. 282) which are embedded in the north German drift and have been derived subglacially from live rock or from preglacially weathered crusts.

Glaciers *v.* Rivers. The difficulty in allotting to glaciers and rivers their respective roles, owing partly to interference by frost, wind, lateral and subglacial streams and other agencies, has led to very contrasted views. Some deem ice to work relatively insignificantly and much more slowly and freebly than streams (see p. 212). Thus Heim,[203] who thought glaciers widened rather than deepened their valleys, estimated their lowering in Switzerland at only a few metres or one-five-hundredth or one-thousandth of what rivers have accomplished. Glacier-muds, as well as the thickness of the drifts (see below), have led others[204] to regard ice as the more potent. Penck,[205] from the Lower Aar Glacier, estimated the ratio in its favour as 2·5:1, an opinion shared by others,[206] including Gilbert[207] who stated that ice worked more quickly in the North American Sierra Nevadas than other agents under similar topographical and structural conditions. Restoration of the initial profiles gave a ratio of 3:1.[208] In the case of the Hintereisferner this was estimated at 10:1[209] and in the Hornafjördur district of Iceland at 5:1.[210] In the Pamirs, valleys below glaciers have a thick carpet of deposits while those without glaciers have none.[211] In Alaska, denudation is faster above ice than in ice-free valleys receiving equal precipitation; freedom from vegetation, due to the cold the ice induces, augments the run-off, while the ice removes avalanched talus and exposes fresh rock to frost.[212]

The erosive superiority of moving ice over running water lies mainly in its rigid hold on material at its base and sides, aided by its power of freezing on to such fragments, especially if there is moisture or tension—an ice-scratch is cut by a grain of sand held in the ice during its short passage from one end of the scratch to the other. Also, while a river cuts a narrow bed or groove, a glacier, which fills its valley from side to side, erodes over a surface, especially in the case of ice-sheets. There are, however, counterbalancing factors. Ice moves much more slowly than a stream carrying equal precipitation from a basin of equal size—its velocity is probably about one-millionth of that of a river.[213] Nor is it liable to periods of flood or great velocity, the times when a river does most of its erosive and transporting work. Moreover, the load which in a river is dragged along its bed is usually distributed throughout a glacier's mass and, unless frozen to the sole, is withdrawn from erosion of the subjacent rock, if not quite from that of the travelling fragments themselves (see p. 382). This is indeed appreciated by those, like A. Baltzer,[214] who deny valley glaciers any but the most insignificant erosion, or by Heim[215] whose estimate of 10 m for the Pleistocene lowering of the Stockholm district is an admission that in these circumstances ice is the more effective.

Rivers are hampered by their low transporting power[216]: an equivalent glacier is able to convey larger masses, both individually and in bulk—its competency in this respect has virtually no limit. Moreover, while water loses this power if its velocity falls below a certain minimum, ice by virtue of its consistency does so only when there is no flow at all: it deposits when it melts, not when its energy of motion is reduced. Nevertheless, a river's

higher velocity, it has been said, more than compensates for its lower power to transport and may in an equal time move very much more than a glacier.[217]

The problem is further complicated by the elimination of chemical action beneath glaciers[218] (sometimes not wholly lacking[219])—Quaternary clays do not contain products of chemical weathering to any noteworthy extent[220] (see p. 226)—and the fortifying of the ice with methods peculiar to itself, such as the ability to pluck, an action almost unknown in rivers.[221]

Faced with these uncertainties, we must confess that our data are inadequate to test the matter quantitatively.

Accurate measurements. In order to remedy this deficiency, at any rate in one respect, steps have been taken to gauge the speed at which ice does erode by measuring the erosion on ground the ice has just vacated or is beginning to invade. Small bores, usually 2 m deep and filled with plaster of Paris or similar material, have been drilled into bedrock on positions accurately surveyed.[222] They have been made, for instance, in front of the Lower and Upper Grindelwald, Hugi and Rhône glaciers, the glaciers d'Argentière and des Bois, Obersulzbachferner, Hintereisferner, Vernagtferner and Guslarferner. Few results have yet been achieved. A cross chiselled 3 mm deep in 1846 has been effaced and the surface striated by an overriding glacier of 1856, as was seen when this receded.[223] The Upper Grindelwald Glacier abraded smooth rocks 0·5–1 mm in 6 months or up to 3·9 cm in *c.* 6 years (1918–24) and plucked out pieces up to 0·2 cu. m in size.[224] The average wear by the Allalin Glacier, mainly by plucking, was 30·1 mm in 5 years or a maximum of 184 mm.[225] On the Biferten Glacier, it averaged 1 cm/annum.[226] But these figures, because the ice was thin and sluggish, help us little. Atmospheric agencies which worked upon the rock before overriding also vitiated them to an unknown degree.[227]

Thickness of drift. E. Collomb[228] was the first to recognise the ground-moraine as a product of ice-erosion: its very existence constitutes proof.[229] Accordingly, if we compute the volume of the drift, we have forthwith a measure of the erosion.

Estimates of the average thickness of the drift, in most cases little more than guesses, have been made in many areas (cf. p. 595). The following are representative: Denmark,[230] 50 m; Mecklenburg,[231] 66 or 50–100 m; North German Plain,[232] 100 m (other averages[233] are 58 m from 470 bores; 51·7 m from 467 bores; not more than 35 m, 75 m or 80 m); north Germany and Russia,[234] 50 m; Norrland,[235] 4–7 m; south of the Tatra,[236] 33 or 50 m; west Edenside[237] (north England), 3–6 m.

Europe's Pleistocene accumulations are exceptionally enormously deep.[238] Thus at Tonning, they are 352 m; near Lübbendorf in Mecklenburg 470 m; near Hamburg, 299 m (unbottomed); near Rostock, 288 m; at Brunsbüttel, 243 m; at Bremen, 231 m (unbottomed); in Jaeren, 124 m; near Grunberg, 154 m; at Lyck, 183·6 m; at Kiel, 167 m; at Strasbourg, 204 m; near Grenoble *c.* 400 m; near Berlin, 201·5 m (unbottomed); at Utrecht, 108 m; at Heidelberg, 397 m; in the Netherlands,[239] over 300 m (the subdrift floor is at −160 m at Beera, at −175 m at Nondlaren, at −200 m at Schoorl, near The Hague at −395 m); in Allgäu 350 m[240]; and in the Romagna much more than 100 m and in the Po valley near Imola *c.* 800 m.[241] In East Anglia,[242] 470 ft (143 m) has been recorded and in the Isle of Man[243] 575 ft (175 m).

The average thickness (in feet) in various parts of North America has been

computed as follows: Minnesota,[244] 100–150; New Jersey,[245] 20–40; Wisconsin,[246] 50–60; Illinois,[247] 100–130; Maine,[248] 30–50; Massachusetts,[249] 15–25; New Hampshire,[250] 10–15 or 20; Connecticut,[251] 5–10; New England,[252] 30–40; over much of the site of Lake Agassiz,[253] 50–200; Coteau des Prairies, South Dakota,[254] 300–700; south of Hudson Bay and James Bay,[255] 100; south part of the peninsula of Michigan,[256] 300; Iowa,[257] 150–200; parts of Ohio,[258] 95·7 (nearly 3000 well records); and elsewhere in that state,[259] 69·5 or less than 56 and generally about 100 over a piece of country 400–500 miles (*c.* 650–800 km) broad.

Exceptional thicknesses, in places suited as traps for deposits, include 530 and 763 ft in Ohio,[260] 660 ft at Toronto,[261] 700 ft in Alaska,[262] more than 1000 ft near Vancouver City,[263] 1080 ft at Watkins, New York,[264] 1250 ft (unbottomed) in the Finger Lakes region,[265] 1100–1300 ft in the Spokane Valley in Idaho and Washington,[266] 2200 ft about the Fraser delta,[267] and 4000 ft at some locality which I have been unable to trace though the figure has appeared for many years in tables purporting to give the maximum thickness of the various geological formations.[268] On the eastern side of the most active geosyncline in North America, Quaternary deposition has been so rapid that fresh cypress wood is found commonly to depths of as much as 1800 ft in south Louisianan oil wells and recent faunas have been reported[269] at depths of 2400 ft and Pleistocene at 3000 ft.

Granted that ice eroded the drifts[270] (see below), the lowering in the various home lands may be computed. L. F. Kämtz,[271] doing this for the first time, estimated the erosion in Scandinavia at 325 m. This was later reduced[272] to *c.* 76 m or enough to fill all the Scandinavian lakes and the Baltic and raise Scandinavia's level by *c.* 25 m. Assuming an average thickness of 30 m for the north European drift, a lowering of 300 m was obtained.[273] Another estimate gave 193 m or, allowing for the drift's porosity compared with the compactness of the parent rock, 154 m.[274]

With these high estimates (130 m for Rügen[275] and 230–270 m for the Swiss Plain[276] are of the same order) we may contrast the following[277]: Sweden, 25 m; Finland, 10–20 m; Scandinavia, a few metres or 25 m. Nansen,[278] assuming the strandflat's height above the sea represents the isostatic uplift that followed the lightening of the land by ice-erosion, obtained a figure of *c.* 10 m—this method was afterwards applied to Novaya Zemlya.[279] The Alpine lowering has been computed at 10–16 m from the drift's average thickness (25 m) over the Molasse country or at 30 m if there be added the quantity probably carried beyond the Swiss borders (A. Penck). This agrees roughly with Neumayr's figure of 40 m[280] or Penck's 36 m for the tract between the Iller and Inn[281] (drift thickness, 60 m) and 13 m for the Isar,[282] though W. Reissinger,[283] from the thickness of the schotter in the foreland of the Iller Glacier, obtained a figure of 350 m for the four glacial epochs, and 430 m for the total Glacial period, including the interglacial epochs.

The different mineral composition of the north German drifts (see p. 939) is correlated with the Pleistocene denudation of Scandinavia: in the early Pleistocene the metamorphic cover, in later times the crystalline rocks themselves were laid under contribution.[284]

Few estimates have been made for North America,[285] because the difficulties are much greater, or for the British Isles; the erosion in the Lake District was stated on inadequate grounds to be 2 m.[286]

The above figures are useful if only to show that the most careful calculations are but crude and uncertain approximations. The errors are fourfold. First, the drift's average thickness is difficult to assess because its upper and lower surfaces are irregular, a difficulty only partially overcome where bores, wells or natural sections, such as stream bluffs or lake- or sea-cliffs, are plentiful. Secondly, material has subsequently been denuded, especially from the older drifts. Thirdly, the ice and its streams carried much detritus into the sea beyond the present coasts. This error is especially serious[287]; for example, the visible drifts in Massachusetts are probably much less, in the Californian Sierra Nevadas, significantly less than the fine powder which the streams removed. This harmonises with observations on Iron Hill, Cumberland, U.S.A.,[288] based upon the iron percentage in the sands of the boulder-train, and the figure of nine-tenths favouring the streams in the Inn valley.[289]

Preglacial residual soils. The fourth and gravest error arises in the opinion of many from preglacial soils incorporated in the drifts. Such weathered rocks are now found beyond the glaciated terrain, as in the "preglacial" river-terraces in the valleys north of the German Mittelgebirge[290] and the Carpathians[291] which contain only local material and warm, freshwater shells.

The depth of the weathered schist in the Blue Range, Viriginia, is 20–50 ft[292] (6–15 m) and of granite in Columbia 80 ft[293] (24 m) while in South Georgia, near Atalanta, it is 95 ft (29 m), in north-west Georgia 200 ft[294] (*c.* 60 m), in the southern Appalachians more than 250 ft[295] (76 m), and in the Klondyke region of Alaska, 60 ft[296] (18 m). Rotting in Brazil is in places 300 ft. (91 m) or 394 ft (120 m) deep[297] and on unglaciated basalt in the German Mittelgebirge 20 m.[298]

The slowly spreading ice gathered up vast quantities of waste which it found ready to hand. The preglacial surface was swathed in soil and subsoil, secularly disintegrated rock and residual clay, produced during aeons of quiet chemical and mechanical subaerial decay. The mantle included landslips, screes, alluvium, sea-beaches and sand-dunes.[299] Even the steep flanks of rejuvenated valleys had their covering.

Although the depth of the rotted accumulations within the glaciated terrain is unknown (it depended upon the nature of the underlying rock, the climate and vegetation, the state of the preglacial drainage and the ratio between weathering and transport), it was probably substantial and approached that just cited for extraglacial regions. The existence of such accumulations is beyond question. They are constantly being discovered,[300] especially in the marginal strip of weak glacial erosion, as in New Brunswick, Nova Scotia, Prince Edward Island, Magdalene Island, New Hampshire, and near the international boundary in Washington and west Canada. Their sheets or isolated patches repose in sheltered situations, e.g. in the lee of crags and in hollows,[301] and with older boulder-clays and gravels in small, drift-filled valleys athwart the ice-flow[302] (see p. 323). American instances,[303] including preglacial placers, have been found in many parts of Canada and the United States. Those from Europe[304] include red clays at the base of the Dutch drifts[305] (weathered older drift?[306]), silts in the buried valleys of the Clyde[307] (cf. p. 1253), weathered gneiss in east Sutherland and Aberdeenshire,[308] ochre on limestone in north-east England,[309] and the infraglacial beach preserved, for instance, in south Ireland where the ice moved off the land[310] (see p. 1252).

Such rotted rock suggests an average erosion for the glaciated part of the United States of 8–9 m.[311]

Preglacial fossiliferous accumulations. Fossiliferous beds of preglacial age within the glaciated regions are much rarer[312]: they have either been removed by glacial erosion or been deeply buried beneath the mantle of drift. They include amber and plants in Denmark and instances from England and North America. These embrace the Saskatschewan gravels of Alberta, Assiniboia and Missouri[313] which are also assigned to glacial (Nebraskan) or interglacial (Yarmouth) times[314]—they are largely composed of quartzite and chert (with subordinate amounts of argillite, limestone, basic volcanic rocks and Cretaceous sandstone and shale) derived from the Rocky Mountains to the west—and the Orange Sand of Illinois, Missouri, Arkansas, Kentucky and Tennessee. This thickens towards the Gulf of Mexico and, though correlated by some with glaciation,[315] is preglacial[316] since it is indurated and ferricreted and passes below the drift within the glacial limits of Wisconsin and Illinois.

The rarity of preglacial floral relics, the few limestone-caves (Dove Holes Cave,[317] Derbyshire, and a cave in Fränkische Schweiz are almost the sole Pliocene mammiliferous caves in Europe) and the complete lack of a molluscan fauna are somewhat puzzling. Many of the remains were caught up and transported by the ice and its streams; there are occasional streaks of brown soil in the drift[318] and pieces of wood or peat in or at the base of the boulder-clays[319] in Sjaelland, Britain and North America (Nebraskan and Kansan ice-sheets). Peats, soils, vegetable mould, freshwater shells and mammals, together with the contents of Pliocene caves, if such existed (these may have been removed by Quaternary denudation[320] or have been absent because the sea was higher or the climate was drier, as postulated on palaeontological and physical grounds[321]) were probably ground to unrecognisable powder by the ice or swept by it or its streams into the sea outside the present lands.[322]

Evidence has been found which suggests that the ice in places advanced over standing and probably living forests in which the annual rings show a marked decrease in the rate of growth only during the last twelve years before death occurred.[323] Nevertheless, the ice may generally have invaded a barren, timberless and storm-swept country. This was induced by the slow climatic deterioration[324] (which attended, as some would say, the uplift preceding glaciation[325]) and the outrush of cold air, consequent upon the ice-sheet's approach[326]—an explanation also offered for the lack of the later part of an interglacial cycle.[327] The chilling which accompanied the expanding ice hindered the growth of pollen grains and kept them small as in later times, e.g. in the Swabian Alb.[328] A seasonal cover of perishable grasses and small shrubs may alone have persisted in the valleys and lowlands where subaerial weathering, solifluxion, heavy rains and swollen rivers strongly accelerated the effects of the cooling, though an enormous expansion of peat bogs and moors may have been a primary result of the climatic worsening.[329]

Drifts and preglacial soils. The loose residuary clay and earth and the preglacial, decomposed rock were scraped off and kneaded together and forced to participate in the glacial movement: they helped to build up the drift. Many geologists, indeed, claim that this is mainly residual[330]; it contains much oxidised material, oddly-shaped boulders, and rounded erratics referable to preglacial streams or talus, as in the case of the Great Barrington

Boulder Train,[331] or the hard nuclei of surface boulders,[332] weathered preglacially or interglacially. Boulders which originated from secular weathering Hobbs[333] has named *saxums*.

While this origin probably holds for the boulders of gossan in the north Carelian drift[334] or those which have one side glaciated and the other deeply oxidised,[335] it can scarcely be true for the vast majority of the rounded boulders in the weathered drift. These, as in the case of the olivine diabase of south Ontario,[336] are still hard and fresh (ice probably plucked the angular and subangular boulders[337]). Moreover, the loose felspar grains are bright and sound and the more readily decomposable minerals (e.g. micas, amphiboles, felspars), as in the North Sea Drift of East Anglia,[338] are abundant and of large size. Roches moutonnées and the sound Scandinavian boulders, scattered profusely and widely over Europe, establish much denudation of bedrock.

The colour and composition of the drift are alike irreconcilable with such an origin. While residual soils are almost entirely robbed of their alkalis and consist of quartz, undecomposed silicates and ferric oxide, glacial clays, as analyses show,[339] have an unusually high calcium and magnesium content and a relatively low potassium-sodium ratio—boulder-clays without lime suggest overriding of weathered interglacial beds.[340] The high percentage of calcium and magnesium carbonates conspicuous in rock-residues[341] implies much mechanical destruction, as does the mineral illite which is one of the main constituents of glacial drifts but being easily altered is not apt to be a constituent of residual deposits.[342] Preglacial soils were also quantitatively inadequate,[343] especially if we remember those transported by ice beyond the present coasts or lost into streams.

The rarity of vegetation in the drift suggests that the preglacial material was carried beyond the limits of glaciation. On a very moderate estimate (ch. L), the Glacial period was sufficiently long to remove the ancient soil completely without assuming a specially rapid transport,[344] particularly in view of the copious waters which poured over the country. The great distances to which distinctive erratics have been conveyed (see p. 124) sustains this conclusion.

While it is doubtful whether much or any residual soil survived in the later drifts of the subcentral areas of prolonged glacial erosion, as in Scotland, Scandinavia or Canada, survival in a remodelled state in the older peripheral drift is admissible.[345] It is said, for example, that the decomposition of this drift is too deep, too uniform and too thorough to be glacial,[346] its boulders are too few compared with the amount of friable material,[347] and it contains much organic material.[348] In some tills, which contain a high proportion of decomposed material picked up by the ice, the decomposed elements which extend through the deposit and not merely in a soil profile occur side by side with fresh grains of minerals readily susceptible to alteration.[349]

This origin is established for the "mixed schotter" in the northern Carpathian valleys where undisturbed debris of preglacial age mantles the interstream watersheds[350]; for the gold placers in some British Columbia drifts[351]; and, less certainly, for the decayed boulders towards the margin of the older drifts on the Columbia Plateau[352] and the gravelly drift in the Lee valley, south Ireland.[353]

The older drifts are composite. They embrace much residual soil which lay in the path of the ice, soft clays and sands from the sea-floor where the

ice invaded the land,[354] e.g. in England (see p. 632), and *Felsenmeere* and frost screes[355] formed in early glacial times and at the onset of each succeeding glaciation. They comprise too the terminal moraines of the advancing ice,[356] extraglacial or interglacial deposits (as included loess concretions show[357]), together with the soft strata of the lands themselves. Thus the source of the English drifts is very largely the great clay divisions, such as the Keuper Marl in Cheshire and the north-west Midlands, Keuper Marl and Lias with Oolite and Carboniferous additions in Lincolnshire and Yorkshire, and Lias, Oxford and Kimmeridge clays and Chalk in the east Midlands and Eastern Counties.

Driftless area of Wisconsin. Yet another line of enquiry compares glaciated with contiguous regions which escaped glaciation. Such an examination was made of the Driftless Area of Wisconsin[358] (see p. 727) enclosed within glaciated country. It has extremely irregular outlines, numberless outliers, unstable erosion remnants, natural bridges, caves, spurs, crags, castellated bluffs, buttes and pinnacles. Its drainage is arborescent with forward grades and its ridges and valleys are not linearly arranged. Yet 1800 measurements show the depth of its residuary mantle averages only 7·08 ft[359] (*c.* 2 m).

In marked contrast, the glaciated frame displays simple outlines of escarpments and a lack of caves and swallow holes in the limestone and of fragile, castellated outliers. Hence, the difference between the two areas arose from the grading up of the depressions and the filing down of the prominences[360]: East Wisconsin has lost *c.* 60 m of weathered and cavernous limestone[361] and has suffered a topographic revolution.

Comparative physiographical studies and deduction. Rapid progress was made towards an appreciation of the quantitative effects we are seeking when observation was directed to compare glaciated with unglaciated lands and when deduction was systematically used to trace out the consequences of the rival theories of protection and destruction. These methods, elaborated by both European and American geologists,[362] in particular by W. M. Davis,[363] have led, as will be seen in succeeding chapters, to a general recognition of forms which viewed together as the glacial *Formenschatz* are insignia of glaciation, i.e. rock-basins described by Ramsay, cirques noted by Helland, U-valleys with their shoulders noticed by Richter and their over-deepening stressed by Penck and Davis, and fjords differentiated by Dana. The increased modification with increasing glaciation, observed in passing from areas of imperfect cirques and troughs to those of huge cirques and pronounced U-valleys, furnishes convincing proof.

Studies of this kind led Penck and Davis, two of the foremost champions of severe glacial erosion, to abandon their earlier opinions[364] (1882) of its effectiveness and caused Brückner to relinquish similar views[365] (1890).

Test cases. The depth of rock removed by Pleistocene ice has been carefully determined in a number of small areas. The results provide useful controls, though of limited application. Some of them show that the erosion was locally inconsiderable. Thus rocks at the foot of steep and high escarpments, facing the ice, in eastern New York State and fault scarps in places in west Greenland are virtually uneroded[366]; the upland south of the Finger Lakes[367] has residually decayed material in place, even on impact sides of scarps, and bear-den moraines composed of preglacially weathered boulders on less steep slopes; solution joints in limestones in central Illinois are

untouched[368]; shales in Puget Sound are incoherent and rotted[369]; coarse, fragmental debris, insecurely held in the rock, is striated[370]; polished surfaces coincide with those of exfoliation[371]; grooves in Michigan have been hollowed out along preglacially weathered joints, and staurolite crystals in its schists, etched out preglacially, still persist[372]; interglacial deposits within the erosional area survive in localities not specially protected, as in Scandinavia[373]; and preglacial karst topography, with solution channels partly intact, occurs on the moutonnée limestone of Switzerland.[374]

Other examples indicate severer but again small erosion. Thus Gilbert[375] found that resistant ledges in the Finger Lakes region had been worn back sufficiently to lose their subaerial contours and had been fashioned into roches moutonnées, miles long and hundreds of feet high. The Niagara Limestone was regular in its outline where it was parallel with the ice-flow but was deeply serrated where it trended at a high angle, its cliff being furrowed back from the escarpment for ½ mile (*c.* 800 m) and to a depth of 9 m: the limestone lost in thickness 3–6 m. Simultaneously, the ice wholly reconstructed the topography of the Medina Shales; it gouged out flutings 12 m deep parallel with its flow, removed old valleys and wrought a new system, lowering the land by 15 m. Gilbert's conclusion that the ice eroded little received later support.[376] It agrees with the relative resistance of the various formations in Iowa to glacial erosion[377] and with the amount of granite in the various moraines of Drammenfjord compared with the granite outcrop.[378]

Rock-trains from small outcrops are useful checks; the Lennoxtown essexite fan (see p. 365) furnishes striking evidence of the importance of glacial erosion in the Glasgow district[379]; the Wisconsin boulder-trains imply a glacial degradation of 15–23 m[380]; and the train from the peridotite of Iron Hill, Cumberland, U.S.A., one of the same order.[381] The average for parts of the English Lake District has been computed at less than 30 m.[382] The size of granite boulders and the altitude of sheet jointing, of preglacial age, in New England granites with respect to present topography make it possible to institute comparisons with preglacial topography. They suggest a removal of at least 3–4·5 m on stoss sides and of 30 m on the lee sides.[383]

Studies of the topography of the Hudson valley however have suggested that the ice planed back the protruding front of the Catskill Mountains with their well-jointed flagstones and shales for a minimum distance of 2 miles (*c.* 3·2 km)[384]—over 1 cu. mile (*c.* 4 cu. km) of rock was removed. Peaks in this area have been half cut away on the south. The giant glacial grooves of north-west Canada (see p. 247) suggest the removal of about a quarter of a cubic mile of rock from an area of 50 sq. miles[385] (*c.* 130 sq. km).

Adverse arguments. The arguments put forward from time to time against any effective ice-erosion are mostly without foundation; they rest on inaccurate observation, on false deduction or on half-truths. To the second or third categories belong statements like the following: ice, especially when thick, is so plastic that it moulds itself upon irregularities on the rock-floor[386] or loses its solidity by pressure-liquefaction (see p. 44); ice, not rock, is eroded[387]; the present curve is not continuous[388] but has rock-barriers, roches moutonnées, inselbergs, spurs, lake-islands and shoals; when once the rock is polished and saliences are lost no opening is lent for attack and erosion ceases,[389] except in unconsolidated strata; plucking is not a real force, since the alleged plucked-masses were often thrown upon the ice from landslides or by frost[390]; thick boulder-clay acts as a protective buffer[391]; and

drumlins, unaltered or only slightly remoulded, persisted when the ice moved across them.[392]

Some of these arguments either overlook the erosive action of subglacial debris—abrasive particles on the underside of a cake of beeswax may scratch a metal surface[393]—or fail to see in unconsumed residuals evidence of glacial incompleteness rather than incompetence. They ignore the fact that new inequalities were created by a "pluck and heal" process[394] (the Upper Grindelwald Glacier tore off pieces even from rock which was partially rounded[395]). Some of the blocks in the drift, striated on one side, probably received their striae while still in place,[396] as is seen in glaciated lands[397] and at the edge of modern glaciers.[398] The large erratics occasionally noticed with deep pot-holes (see p. 238) have the same significance.[399] To interpret the rough lee side of roches moutonnées as untouched preglacial surfaces[400] is likewise to disregard plucking since they are unweathered[401] and pieces frequently occur which, though some distance away, match their scars exactly.[402] Moreover, rocks are often less solid and free from joints in the lee of crags than on the impact side—in constructing the C.P.R. of Canada, it was found more difficult to remove the rock on the northern (impact) than on the southern (lee) side of bosses.[403]

Drumlins once formed are highly persistent since their adhesive clays, free from joints and with rounded form, are very resistant to both plucking and abrasion.

Great erosive power may be reconciled with great boulder-clay fabricating powers by confining the two essentially opposing forces to different regions or to different periods.

The frequent appeal to cross striae as proof of the feebleness of ice-erosion[404] assumes that they were engraved at widely separate times—which may not have been the case (see p. 902)—or that boulder-clay had not protected the rock during the interval.[405]

It may, however, be granted that when plucking has reached the depth of tight and widely spaced joints its action will virtually cease (see p. 250) and that with the abrasion of this plucked surface erosion will be at a standstill: the ice would then become protective. Greenland and the Antarctic may alike have reached this state.[406]

Glacial cycle. Existing forms of glaciated mountains vary not only with the preglacial stage of development but with the competence of glaciers to erode and with the duration and intensity of glaciation. Davis[407] extended the points of resemblance between rivers and glaciers into the ideal cycle of glacial denudation with its elements of time, rock-resistance and velocity and volume of the ice. The term cryoplanation has been given to this action of ice and frost.[408] Barriers and basins are the respective glacier-equivalents of broken rapids and steadily flowing reaches and alike characterise youth. As the cycle advances, steps are worn back, lowered and eventually destroyed,[409] and as maturity passes into old age, sides are smoothed, floors are graded and cirques and similar forms are eliminated.

The concept of a cycle has found much favour[410]; the plateau at 250–350 m in Novaya Zemlya has been regarded as an end-product of ice-sheet erosion,[411] and similar levelling has been postulated[412] for central Norway, for the Ural Mountains, for parts of Greenland and for the Laurentian peneplain. In the ideal case of a long continuing glaciation in a suitable climate, erosion would be expected to proceed to its completion, though the final form would be accidented

by a small-scale relief of irregularly scattered, steep-sided knobs and mammilations, unrelated to any base-level.[413] Yet it is more than doubtful whether the cycle is realised, unless the glacial peneplain has descended, as in Labrador and Russian Altai, from an earlier, stream-formed peneplain.[414] Not only is glacial erosion not bound to a profile of equilibrium (W. Penck,[415] however, thought this existed at every snout, whether supermarine or submarine) but the Glacial period, minus its interglacial epochs, was too short for maturity to be attained (see p. 918). The end of the cycle is nowhere discernible; the youthful stage only is seen with its inherited features accentuated. Nivation and cirque recession, for instance, are still in an early stage. Glacial troughs are unfinished; they have unconsumed spurs on their sides and repeated steps and basins on their floors. Moreover, prolonged erosion, it has been argued, would exterminate the glacier long before the cycle's final stage.[416] Nivation (see p. 303) also probably fails to achieve a complete cycle,[417] as does the action of solifluxion (see p. 568).

Ice-sheet and glacier erosion. It is essential to distinguish between ice-sheet erosion and glacier-erosion.[418] While glaciers, acting linearly, accentuate relief, ice-sheets, which move more freely over comparatively even ground without the valley constrictions that accelerate flow, pare down the elevations and soften the outlines. By their planar action they work more or less evenly on surfaces which are flat or of gentle relief,[419] as on the high plateaux[420] of Fennoscandia, Iceland, Scotland, Altai and east Greenland, and the strandflat and the low coastal peneplains[421] of Alaska and the British Isles (Lewis, Anglesey, Donegal)—the rounded form, the "Nut", abounds in south Norway[422] and the Finnish lakes are in general surprisingly shallow.[423] They accomplish little because they move slowly, are not concentrated along definite lines, and have little debris. The central Swedish peneplain was probably lowered only a few metres,[424] the Finnish peneplain a few tens of metres,[425] the Stockholm area 10–25 m,[426] and the Canadian Shield a few tens of feet at the most.[427] The Greenland ice-sheet has shielded the underlying peneplain from the time of its regional uplift[428] and the "Gondwana" land-surface in Antarctica may have been similarly protected.[429] Concentration into glaciers on the Norwegian side and ice-sheet planar action on the Swedish side explain why Scandinavian fjords are restricted to Norway and occur generally elsewhere on western coasts (see p. 346) which for other reasons were deeply revived preglacially. The greater velocity in the valleys tended to concentrate the drift-laden ice along these lines.[430] On the uplands between the Finger Lakes, where the ice was thin and its basal load scanty, the preglacial weathered mantle is still preserved in places.[431]

While valley glaciers performed "directive erosion", ice flowing over broad surfaces effected "selective erosion",[432] though to a much less degree than running water. It worked along lines of jointing, bedding, cleavage and fracture, or along weak or soft rocks.[433] Nevertheless, it was not always more competent in soft and loose material. Thus gravels and similar materials caused overloading and stagnation[434] and were only removed by abrasion grain by grain,[435] unless their interstitial water was frozen and conterminous with the ice; the gravel then moved as part of the glacier.[436] For instance, the smooth Eo-cambrian clays between Viborg and Leningrad and the Cambrian clays in Estonia were overridden without being disturbed since they offered no hold.[437]

Date of ice-erosion. Glaciers operated quite differently at different periods of their history. Their erosion often dates from the early phase and decreased as the surface became smooth (see above), asperities were removed,[438] and preglacially disintegrated material and loose blocks were carried away[439]; the largest blocks are in the lowest drifts.[440] The augmented debris slackened the flow,[441] while the higher plasticity (which lessened the hold on the englacial debris) resulting from the augmented ice-thickness[442] and the increasing depth of the protective subglacial drift[443] (which served as a lubricant) also lessened erosion. Yet erosion was by no means confined to such early stages since it was being performed at all times, even during the retreat (see p. 308) whenever conditions permitted.

Conclusion. Investigations and discoveries during the last fifty or more years which have greatly advanced our knowledge and eliminated many erroneous views have established the quality of ice-erosion and to a less degree its quantity. A vast body of well-attested facts proves that the truth lies between the extremes of glacial ornamentation and glacial sculpture though the amount was in one place considerable, in another negligible. Ice scoured off preglacial residual soils and rotted rock over the dispersal zones and sheared off much live rock, rounding ridges, softening asperities, and effacing surface-inequalities in the process. In its positive aspect, its sculpture transcended rock-structure. It developed certain peculiarly typical features, such as rock-basins, cirques, U-valleys and faceted spurs, and accentuated others which were inherited, such as steps and treads on valley floors. Its negative aspect is seen in the modification or obliteration of topographical forms typical of water-action and in the present palimpsest scenery of glaciated lands. It remodelled rather than created.

Classification. The discussion of the glacial erosive processes and their morphological results which occupies the remainder of this part of the book will proceed in the order of the following classification:

Erosion within the Area of the Ice

1. By subglacial streams
   - (i) giant kettles
   - (ii) subglacial canyons
2. By ice
   - (i) Minor features
     - (*a*) abrasion
     - (*b*) plucking
     - (*c*) roches moutonnées
     - (*d*) crags and tails
     - (*e*) subglacial disturbances
   - (ii) Major features
     - (*a*) rock-basins
     - (*b*) cirques
     - (*c*) U-valleys
     - (*d*) Fjords, fjärds, föhrdes

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CHAPTER X

EROSION BY SUBGLACIAL STREAMS

Subglacial streams are apt to be neglected or underestimated in discussing the larger question of glacial erosion. Yet we may gauge their importance from the vast bulk of the fluvioglacial silts which amount, for example, to millions of cubic metres in the Swedish varves,[1] from the enormous volume of the Swedish osar, or from the insignificance of the moraines and coarse sediments compared with the valleys and hollows which have been glacially eroded.[2] They have indeed been regarded as the prime cause of ice-erosion.[3] Their erosive features are giant cauldrons and canyons.

1. *Giant Kettles*

Morphology. Giant kettles (Ger. *Riesentöpfe*; Fr. *marmites*; Swed. *Jättegrytor*; Norw. *Jaettegryder*) are pot-holes or cauldrons which subglacial streams excavate in rock or drift. F. Svenonius[4] fully discussed their shape. Their ground plan is more or less circular or elliptical, the major axis trending with the striae and ice-flow or with the joints along which they lie as deep, narrow clefts. A sinuous channel from the side sometimes betrays the stream's course.[5] The walls are smooth and either vertical or aslant. A water-worn groove, formed by the water's oblique entry, often descends the kettle spirally. Immature types are funnel-shaped while wider and deeper ones expand at their base. Alpine kettles were undercut on the upstream side; the higher ice flowed more rapidly and the moulins were oblique.[6]

Depth like the breadth is very variable and is limited by the hardness of the rock, the volume of water and time. It may be 6 m in Sweden[7] and 15 m in North America.[8] The ratio to breadth is 5:1 in Scandinavia and Minnesota.[9]

Embryonic kettles[10] are oval or round in outline. These small depressions may be the only parts of kettles which have survived; for kettles commonly have sharp and abrupt edges or an ice-moulding on the rims,[11] and some have been wholly removed and occur on erratics[12] (see p. 229). Other imperfect varieties, presenting half a cylinder, may be seen on steep rock-faces[13]: the complimentary half was in the glacier.

Distribution. Giant kettles are plentiful in almost all glaciated regions, as may be gathered from G. Leonard's summary[14] of those found before the middle of the last century. They are especially numerous in Fennoscandia[15] (cf. Finnish list[16]) but are rare outside this central area, being sporadic over Germany,[17] including the celebrated ones on the Muschelkalk at Rüdersdorf,[18] in Holland,[19] and in the Baltic provinces.[20] The Vosges[21] and Pyrenees[22] have few and the Ostrobothnian plain apparently none.[23] Their scarcity in Switzerland is brought out by Heim's list[24]; they include the nine (of Bühl age) in Tertiary sandstone in the well-known Glacier Garden of Lucerne,[25] produced by the Reuss Glacier; these nine, each an annual product,[26] have a maximum depth of *c.* 8·5 m and width of 8 m and an occasional spiral.

British instances are surprisingly few,[27] possibly because they coincide with the thickest drifts and with the present drainage. They have been discovered in Mull, near Cambus o'May (Deeside), in Glen Nevis, at Birkenhead (the Kailpot), near Grasmere, on the ridge between Rydal and Grasmere, near Ilkley and at Ballylusk, Co. Wicklow.

Others have been found in Kerguelen,[28] Iceland,[29] west Greenland[30] and Baffin Land,[31] and more abundantly in Canada[32] and, widely scattered, in the United States.[33] Thousands of pot-holes and cuspate channels, aligned in groups, pit the eastern flanks of the Front Range of Colorado,[34] the largest being 61 m by 46 m and 15 m deep.

Origin. Giant kettles were early ascribed to the activities of giants, as in Norway, or of aboriginal Indians in America. They were afterwards attributed to the weathering out of stone kernels, to vortices,[35] to whirlpools in the *petridelauniske Flod* (see p. 617), or to the excavating power of ice.[36] Charpentier[37] was apparently the first (in 1841) to connect them with glacier moulins. Not only have the latter been occasionally observed at work under modern glaciers[38] but kettles, whether single or grouped, are unrelated to modern streams and often occur where such streams could not possibly erode them. Moreover, they are frequently arranged in rows parallel with the ice-flow (*Paternostertöpfe*), as given by striae or osar,[39] or are situated just where crevasses would arise,[40] namely in the lee of steep faces and roches moutonnées, above steep falls and hill sides, on the edges of barriers or Inselberge, and of projections generally, as so commonly in Finland.

The mere impact of falling water was probably not enough to excavate the cauldrons. This was generally done by swirling stones and boulders or by sands and gravels.[41] The elliptical eroders are sometimes still retained, though more often they have been swept into the drift, some of whose spiral or elliptical boulders may have originated in this way.[42] That kettles usually occur singly may be because the moulin was stationary or its upstream side melted back as the ice flowed on.[43] They probably belong to the later stages of glaciation[44] (not early, as occasionally postulated[45]) when the ice was thin and melting rapidly and moulin shafts reached its base: they are eroded into striated surfaces and their overflows are unstriated.

Not all kettles arose at the bottom of moulins. Some, as Sefström[46] recognised, were the work of modern streams (as has been suggested for the Glacier Garden, Lucerne[47]) and others were hewn by the waves of epiglacial seas[48] which fashioned mixed forms if they had access to glacial kettles.[49] Many were hollowed out by subglacial streams[50]; they often lack, for instance, the elongation which the moulin hypothesis requires; they came into existence after the ice had become stagnant, as their unglaciated rims testify; and some arose at depths well below the then level of the sea, e.g. in coastal Scandinavia,[51] though some in north Norway at least may have originated when the land was higher.[52]

Other agencies may simulate giant kettles. In Germany, for example, solution pipes in limestone have been so interpreted (ice may have modified some like those from Lago d'Iseo[53]), while depressions hollowed out of the granite of the Riesengebirge[54] by forces at present in action, but erroneously identified with subglacial kettles, led to the belief that these mountains had been extensively glaciated.[55] "Weather pits", as in the Sierra Nevadas and Rocky Mountains of North America, may also resemble giant kettles.[56]

It has been held that rock-basins, like Loch Morar[57] or others in north Germany or the English Lake District,[58] cirque-lakes,[59] or even cirques themselves[60] originated by moulin action and by uniting glacial pot-holes as seen north of Lake Athabasca.[61] The cumulative effect of moulins working up and down a glacier has been emphasised too from Himalayan observations.[62] E. Geinitz,[63] who stressed this action, named lakes so eroded "eversion lakes". But this origin is inadmissible[64] in the case of cirque-lakes as well as in all other lakes except the tiniest since the moulin's range, especially laterally, is far too small.

2. *Subglacial Canyons*

Subglacial canyons are not so well known as giant kettles though many instances have been described,[65] e.g. from Norway (interpreted as interglacial subaerial river channels[66]), Baltoscandia, north German Fläming, Adamello, Durance valley and North America. They are frequently buried under retreat moraines or later drifts (as giant kettles were occasionally[67]) and have only been revealed when these have been stripped off artificially or naturally. Such discoveries suggest that canyons are common[68] and that the drainage was deranged while the ice lasted as well as during the coming and going of the ice (see ch. XXIII). They were excavated by the great hydrostatic pressure of the subglacial streams,[69] especially in connexion with stagnant ice[70] or in soft rocks, or in hard rocks where the supply of rock-fragments was adequate. They are usually narrow and deep and water-worn throughout unless subsequently ice-moulded (see p. 113).

The difficulty of distinguishing them from hollows made in other ways is considerable; for though a subglacial origin may be suspected[71] where a barrier is notched laterally or has no lake-terraces above it, they may have been eroded superglacially as spillways (see ch. XXIII) or extraglacially by fluvioglacial streams,[72] as suggested for the channels that stripe the plain outside the Russian Altai[73] and for the channelled scabland netlike system of "coulees", occupying at least 2000 sq. miles (*c.* 5000 sq. km), on the bare basalts of the Columbia Plateau, Washington.[74] These begin abruptly at the limit of glacial advance of pre-Wisconsin age, a few miles south of the Spokane and Columbia rivers. These Columbian channels, up to *c.* 300 m deep, have, however, also been attributed to direct glacial erosion by a scabland lobe,[75] to overflows from glacier-lakes ponded in the Columbia canyon (see p. 459), to ice-jams of enormous size originating in the gorge,[76] or to melt-water streams of normal discharge from glaciers which overran the northern part of the tract and scoured the basalt along preglacial drainage lines.[77]

Canyons may also have been wholly or partly excavated by postglacial streams,[78] as observations on the Upper Grindelwald Glacier or the Kennicott Glacier, Alaska, suggest.[79] Some of the Swiss gorges,[80] e.g. the Kirchet above Meiringen on the Aare, the gorge of St. Moritz on the Rhône, that at Rosenlaui at the Lower Grindelwald Glacier, and others at the end of the Aletsch and Mer de Glace, all usually regarded as postglacial, may be partly subglacial since subglacial waters created new valleys and widened old ones.[81] For example, the Aare gorge has been filled and replaced by a new one five times.[82] Multiple notches are known elsewhere.[83]

Subglacial streams have been invoked for the sublacustrine channels in Lake Geneva and Lake Constance[84]; for the excavation of rock-basins,[85]

including the subalpine lakes; and for the deepening of U-valleys (see p. 330)—their share has been estimated[86] at 5% (W. M. Davis), 20–25% (J. Brunhes) or 60% (W. Kilian).

Rinnentäler. Although such action, except in a minor degree, is improbable in connexion with the features just mentioned, its role may have been not inconsiderable in the *Rinnentäler* and their *Rinnenseen* (Dan. *Langsöer*) or furrow lakes. These long, narrow, river-like channels, with flat or trough-shaped floors (fig. 46) and strings of lakes, are usually flanked by sharp erosion profiles. The lakes themselves are generally long and narrow[87]

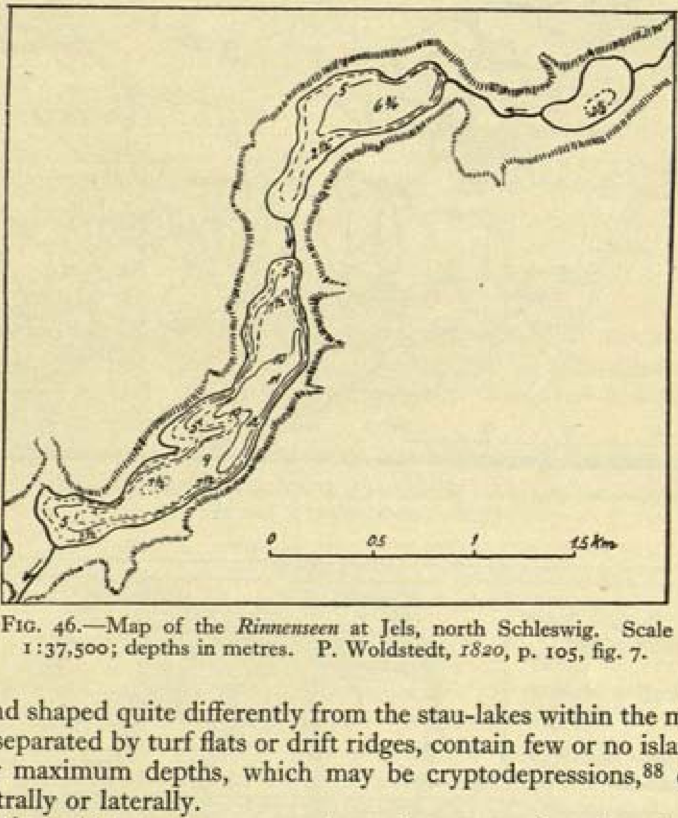

FIG. 46.—Map of the *Rinnenseen* at Jels, north Schleswig. Scale 1:37,500; depths in metres. P. Woldstedt, *1820*, p. 105, fig. 7.

(fig. 47) and shaped quite differently from the stau-lakes within the moraines. They are separated by turf flats or drift ridges, contain few or no islands, and have their maximum depths, which may be cryptodepressions,[88] disposed either centrally or laterally.

Rinnentäler, which traverse ground-moraine, outwash and end-moraine, are parallel with the osar and transverse to the end-moraines. If these are lobar, as around the *Zungenbecken* of the Narew depression, they are radially arranged[89] (fig. 48) but pursue parallel courses where the ice was unable to deploy.[90] The ice-lobes in Denmark have hollowed out their beds into tongue-like basins.[91]

Rinnenseen are widely distributed near the margin of the Baltic moraine,[92] notably in the outwash terrain between the Pomeranian and Brandenburg phases, in the Baltic lands,[93] in north-west Russia[94] and as the "tunnel-valleys" of Denmark[95] which are up to 75 km long—much shorter ones occur in the Danish islands. The *Rinnenseen* attain lengths of up to 17 km or, as strings of lakes, 60 km.[96] Most north German lakes fall within this category.[97] Similar *Rinnenseen* are associated in north Germany with still earlier phases,[98]

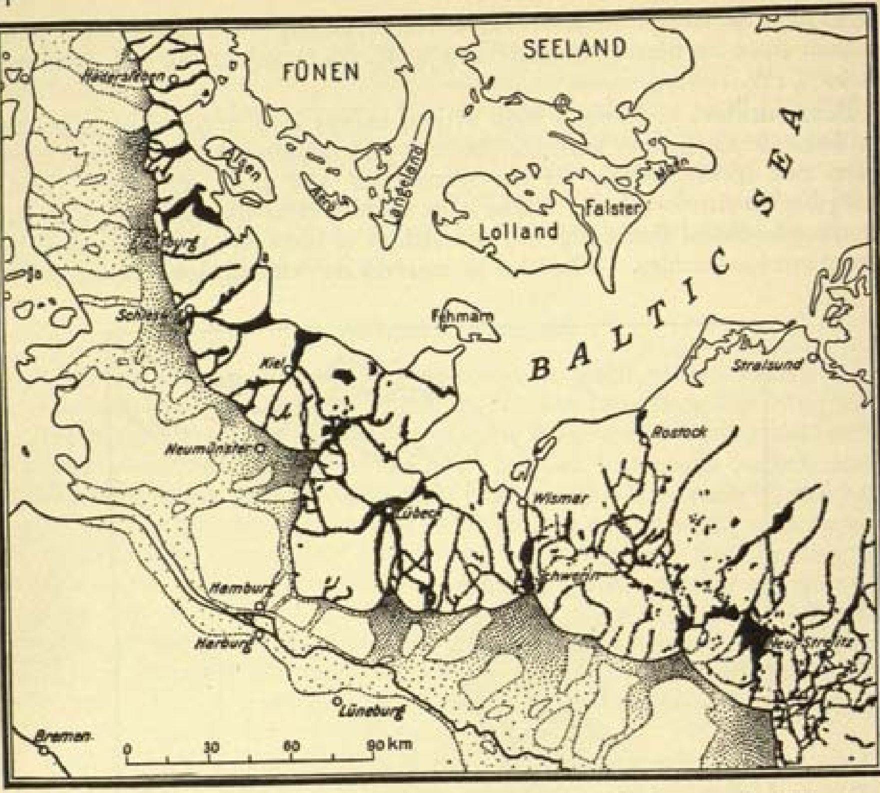

FIG. 47.—*Rinnenseen* and cones in Schleswig-Holstein and Mecklenburg at the Baltic stage. P. Woldstedt, *1820*, p. 120, fig. 12.

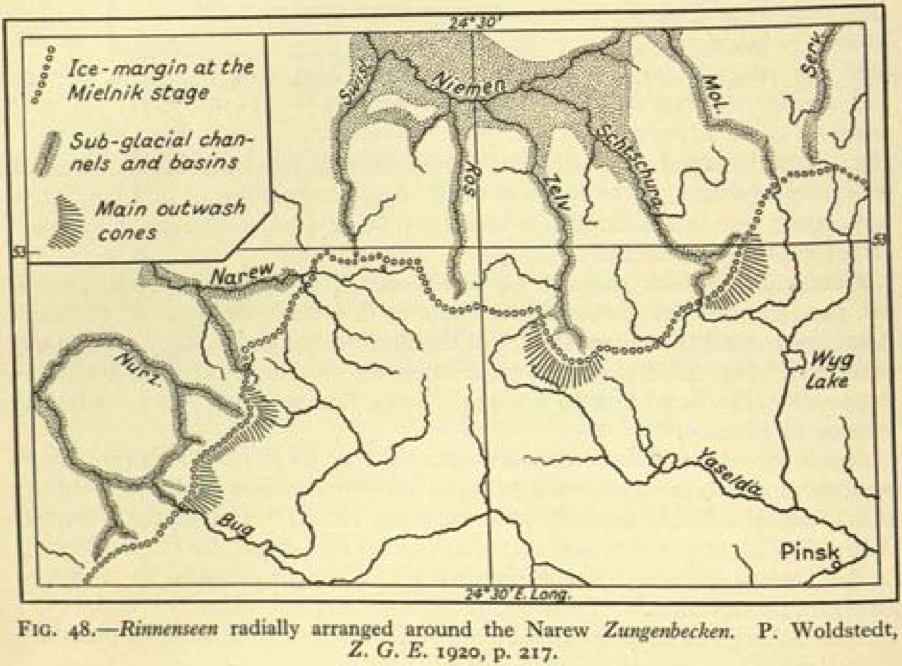

FIG. 48.—*Rinnenseen* radially arranged around the Narew *Zungenbecken.* P. Woldstedt, *Z. G. E.* 1920, p. 217.

as in Silesia and the Lüneburger Heide, and occur on Pre-Cambrian rocks in Sweden, possibly on the floor of the Gulf of Bothnia,[99] and on the Schotter Nagelfluh and Molasse of the Alpine Foreland.[100] The *Kursudalar* in Norrbotten (north Sweden) also seemingly belong here[101]: they are dry valleys, up to 9 km long, with a series of small but deep lakes separated by rock-bars. Some of them are related to osar.

Since Berendt[102] first recognised the type, they have been generally related to glacial streams,[103] either extraglacially[104] (the hollows coincide with colks or eddies[105]) or, more commonly, subglacially,[106] guided by crevasses.[107] This last view is borne out by their parallelism with the ice-flow, as Berendt noted, and their arrangement perpendicular to the ice-margin, whether this be straight or lobar. Moreover, they are associated with osar[108] and outwash fans (see pp. 427, 438); for where they are imperfectly developed, outwash fans are similarly poor. Some writers, indeed, equate the lakes with giant kettles,[109] their valleys with retreating moulins,[110] and their sharp bends with crevasses.[111] Their section, which seems to rule out this origin for most of the lakes,[112] has led many to ascribe them to moving ice[113] (possibly by the widening of subglacial stream channels) and the retention of their shape to inert ice[114]—the Norfolk Broads were similarly ascribed to a late occupation by surface ice.[115] Some *Zungenbecken* (see p. 269) are deemed to have been interglacial stream channels widened by ice.[116] As the cones outside the ice were gradually being built up (see p. 438), the subglacial streams became englacial and finally even superglacial, flowing over ice that filled the lower parts of the channels[117]; rising subglacial streams would not be able to build the outwash delta cones[118] (see p. 437).

Woldstedt,[119] who was of the opinion that ice-oscillations converted some *Rinnenseen* into *Zungenbecken*, such as the föhrdes (see p. 353), thought each depression in a *Rinnensee* marked a recessional stage and was preserved from infilling with sediment by dead ice which lingered because the ground was frozen. The constancy of the lobes fixed the crevassed zones along the planes of contact and so encouraged streams to excavate along well-marked lines and channels.

The objection that such streams would erode the softer ice and not the rocks[120] is invalid, as is the attempt, in virtue of their general parallelism, to associate *Rinnenseen* with tectonic lines[121] (in north Germany, Jutland and Finland). This hypothesis fails to relate the *Rinnenseen* to joint or fault systems or to acknowledge that they are perpendicular to the ice-margin.

Recently, the *Rinnenseen* have been dissociated from subglacial streams and related to crevasses, so that drift was deposited in the dirt-laden ice on either side but not in the ice itself.[122]

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Chapter XI

MINOR EROSIVE FEATURES

1 *Abrasion*

Striae and grooves. Ice acts as an enormous flexible rasp. The abrasion, which varies with a rock's porosity, cohesion, hardness and structure, is severest where the relief offers most resistance and concentrates the flow into valleys.[1] It arises from basal friction which is linked with the depth and velocity of the ice, the quantity of the basal debris and with the nature of the contact and degree of adherence of the ice to its bed. The friction is not even approximately known though it has been sought experimentally[2] and by applying McConnell's coefficient of plasticity to observations on the Hintereisferner.[3] The reduction of the flow at the sole is, as J. Tyndall[4] recognised, a measure of the erosive power. This is a function of the weight and downstream impulse of the ice and may vary less rapidly than their product.[5]

Chamberlin[6] in 1888, in a masterly analysis of rock-scorings, fully discussed the origin and significance of the various types: his interpretations remain essentially those of modern glaciology.

Markings inscribed by ice, shod with basal debris fixed like the teeth of a file, vary in their coarseness with the nature of the rock as Sefström[7] noticed: they constitute part of his "friction phenomena". They range from hair-fine lines which to be seen require to be moistened or rubbed with fine clay or graphite or viewed in a low sunset light, to grooves, many feet wide and hundreds of feet long, which simulate the flutings on a Doric column. Finer striae characterise tough and fine-grained rocks, like quartzites or quartz on which sometimes alone they are now preserved, or on soft rocks, e.g. limestone, serpentine or clay slate, which are apt to be minutely and delicately striated. Coarse rocks, such as grit, take on only the bigger and deeper corrugations.

"Pseudoglacial scratches"[8] may be made by other agencies: these include buffaloes or other animals,[9] boulders trailed by seaweed,[10] mountain torrents,[11] winds[12] and landslips,[13] partially consolidated lavas gliding over each other,[14] creep, mudflows or solifluxion,[15] sliding snows,[16] tectonic movements[17] (with faceted boulders[18]), as well as avalanches (see p. 579), river-ice (see p. 578) and drift-ice (see p. 587). Striae, however, are typical memorials of land-ice; for as E. Desor[19] said long ago *le grand argument en faveur de la théorie glaciale est et sera toujours la roche striée*. Striations made by torrents are short and more or less curved[20] while those made by creep and solifluxion are usually more superficial and mainly found on softer and angular fragments[21]; they rarely occur on two faces and run down and not along hillsides as glacial striae often do. Tectonic striations seldom fall into a system and when engraved on boulders tend to follow the curvature.[22]

Glacial striae, as demonstrated experimentally[23] and observed on modern Alpine glaciers,[24] are scored by boulders, pebbles, gravel and grains of sand frozen into the sole and dragged along between ice and bedrock. Their

character depends upon the nature and shape of the scratching tool (soft rocks may scratch hard ones in experiments[25]), upon the force, steadiness and firmness with which it is driven, and upon the texture and resistance of the two friction surfaces. The finest scratches were grated by fine sand, probably not gripped in the ice but rubbed between large erratics and the bed. They ceased when the grain was liberated. Shallow and narrow at first, they steadily deepen and broaden to end abruptly or to decline again in depth and width and die away. The ice moved towards the broad end[26] as it ground away the point of the inscribing instrument, though the latter was sometimes rotated or wrenched out of position where the scoring terminates in a hole. The law is neither universal nor altogether trustworthy[27] (the reverse has been said to hold on downward slopes[28]); for the marks may commence abruptly and taper away gradually if the boring tool came suddenly into action and wore itself away gradually or was withdrawn. Teeth on jagged striae point in the direction of flow.[29]

The high proportion of material, including boulders in the sole, caused continuous striation by preventing retraction into the ice.

Round or nearly equidimensional boulders are apt to rotate on encountering an obstruction. Striae which end sharply sometimes begin again a little to one side of the original line after an interval that is bridged by irregular abrasions made by a pebble rolling over and resuming its work.[30] Rotation is also proved by the curvature and shortness of the striae and by an occasional screw-like furrowing on the flanks of groves.[31] Screwed or curved striae, early ascribed to floating ice swinging when partly aground, arise when boulders turn in the sole, as has been seen for example on the side of the Gorner Glacier.[32] That fragments escaped from the grip of the ice is proved by the forms just mentioned and by others to be described, including chatter-marks, crescentic gouges and boulder facets.

Grooves, apart from moulded preglacial hollows or subglacial stream-channels (see p. 113), are, as Sefström[33] noticed, made by boulders, notably those of quartz, quartzite, porphyry or other hard material. Their depth, which is proportionate to the pressure of the scorer and the softness of the subjacent rock, may diminish as the tool wears away or may deepen as it is ground into the rock until it stops, jumps out or turns over. Some were eroded by dirty streaks within the ice,[34] others by pebbles or boulders packed tightly together and propelled as a single mass, fluting the furrow's sides. Those with uniform cross-section were scoured out by a single tool, broader and less regular ones by boulders travelling in succession along the track.[35] Thinner, parallel striae replaced a strong furrow when the scorer was crushed into fragments. Grooves, of unusually large size and up to 30 m deep and 1 mile (*c.* 1·6 km) in length, have been observed in rock in north-west Canada[36]—others occur in till (see p. 389).

Great friction is seen not only in the bruised and rough edges which striae often show and in the occasional evidence of a slow rocking motion, but in the system of fine cracks running at a slight depth under the striated surface when examined in thin transverse section.[37]

A glass-like polish, which may be produced by faulting ("fault mirror"), sand drift,[38] ice and rock-particles[39] or by animals,[40] results generally from glaciation and is most perfect, as Agassiz[41] observed, on hard and fine-grained rocks like quartz and quartzite and on their boulders in conglomerates. It is imparted by the friction of smooth flat surfaces of boulders frozen firmly into

the sole, by the fine powder derived from preglacial clays and soils or the great clay formations, or by the finest matrix of the ground-moraine or rock-flour ground by the ice. It implies not a little erosion, possibly twice that of striation[42] though this bears striking testimony to the severity of ice erosion.[43]

Time of formation. Striae may have been inscribed at any time but, unless towards the periphery, probably during the later phases.[44] This age, anticipated by Agassiz[45] (on the false premise that an ice-sheet because of its low temperature was immobile except at that time), is supported by the striation's common orientation perpendicular to the ice-margin ("submarginal striae"[46]). This relationship, though not invariably true where the ice-borders were subaquatic and were much influenced by fracture,[47] has been shown, for instance, in Greenland[48] and for the lobes about the Great Lakes.[49] Geochronological studies and the crescent-shaped *Salpausselkä* reveal that the receding ice-edge in Fennoscandia[50] crossed the striae.

Associated markings. Chattered striae or "chatter marks"[51] (E. Collomb's *saccades*[52]) are curved, transverse lines or fractures, arranged along an axis. Found in firm but brittle rocks, e.g. granite, basalt or quartzite, they

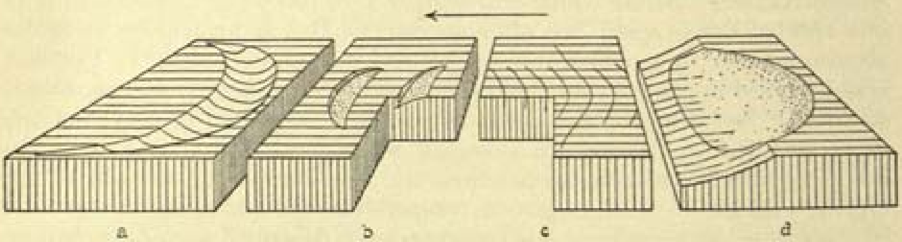


FIG. 49.—The various marks produced by ice on rock-surfaces: (*a*) *Sichelwanne*, (*b*) crescentic gouges, (*c*) crescentic cross fractures, (*d*) conchoidal fractures. E. Ljungner, *1919*, p. 287, fig. 123.

are so minute and densely packed that often only a close inspection detects them. Appearing as a succession of bruises, they were made by a breaking out of thin rock-surfaces through compression in front and tension in the rear of a point of application. The boulders, partially and insecurely embedded in the ice, moved vibratingly and unsteadily and struck with slow rhythmic or jerky effect.[53] Jagged grooves, which roughly fracture the rocks, are related forms. They are alike the result of more intense action of the ice and higher bearing pressures which momentarily and locally reach the rupturing strength of the rocks.

Cross fractures[54] are steeply inclined cracks, convex upstream: "crescentic cracks",[55] "serrated striae"[56] and *Parabelrisse*[57] are identical or similar. They are generally smaller (less than 7·5 cm) and rarer than crescentic gouges[58] and almost confined to finer grained rocks. Experiments prove that they were made by the gouging corner of a boulder held in the ice which set up shearing stress.[59]

Crescentic or jumping gouges[60] (fig. 49), which are also included in these "friction cracks"[61] on stoss sides, are concave markings up to 15 cm long which occur in sets aligned on a common axis and with the flow, the members of a set conforming to a particular size or enlarging towards the axis or in the direction of motion. Associated with striae and grooves on the impact side

of obstructions or near the crest of hills of hard, massive, homogenous or brittle rocks, they result if the ice froze on to the rock[62] or if friction was increased as by sand pockets in the sole.[63] Alternatively, they arose if a block (E. Ljungner) or pointed or edged tool (F. H. Lahee) oscillated as it moved obliquely forward upon a polished surface. It raised a wave in the rock in front of the point of application, failure taking place along a portion of a cone whose axis was inclined forward.[64] Hence crescentic gouges are closely related to chatter marks but differ from them in that the chiselling body, acting only at intervals, skipped over intermediate distances.

Lunoid furrows, forms equivalent to crescentic gouges,[65] are *c.* 30 cm long between the horns and abrupt on the concave or iceward side. They are most frequent on impact sides, occurring as solitary lunes but usually in groups aligned with the neighbouring striae. R. A. Daly ascribed them to shearing stress set up by the dragging of a subglacial boulder, the tension being relieved by initiating incipient cracks dipping gently forward. Their occurrence in widely contrasted rocks suggests that structure is quite subordinate. The actual lenses, in accord with the freshness of the furrows and the absence of covering lichens, were possibly split by postglacial frost action and were limited in size by the distance measured along the crack through which the rock was rendered unsound. Vertical pressure of a boulder flaked out a circular or elliptical disc or "annular scar".[66]

Conchoidal fractures,[67] which have been observed in process of formation,[68] are independent of structure and occur where a rock-surface falls away.

Terminal curvature. The severity of glacial pressure is seen not only in the features just described but in the twisting back of the ends of the more yielding strata as first noticed in several places in Britain in the middle of the last century[69] and more recently in that and other countries.[70] Such bending may be due to other causes,[71] such as solution or faulting, but is mostly referable to creep or solifluxion (see p. 567) as in unglaciated Cornwall and Devon.[72] It may, however, be made by the direct impact of ice, as by the present Vernagtferner,[73] but can only be invoked with certainty if the bending has been uphill.

2. *Plucking*

Plucking or the detachment of rock-fragments by overriding ice is facilitated by structural planes,[74] e.g. stratification, joints, cleavage or foliation (fig. 50). Thus the boundary between abraded and plucked surfaces is sometimes that between different kinds of rock. The action is especially powerful if the planes make a small angle with the surface or emerge upstream, or if the strata are thinly bedded or jointed or broken, as along shatterbelts or at fault intersections.[75] Strongly jointed igneous rocks have a marked disproportion of boulders in the drifts.[76] This has repeatedly been observed in the case of granite,[77] especially if the rock had previously been subject to exfoliation, as in British Columbia.[78] In the Lake District, the hard and well-jointed Borrowdale rocks have been severely plucked and have yielded numerous large erratics while the Skiddaw Slates with their innumerable closely placed joints were removed in small fragments and produced a tenaceous till.[79]

Plucking was severe if support was weak, as on boulders in conglomerates, downstream from projections like roches moutonnées (see below), lateral spurs, valley steps, diffluence divides and the sides of subglacial streams, or where

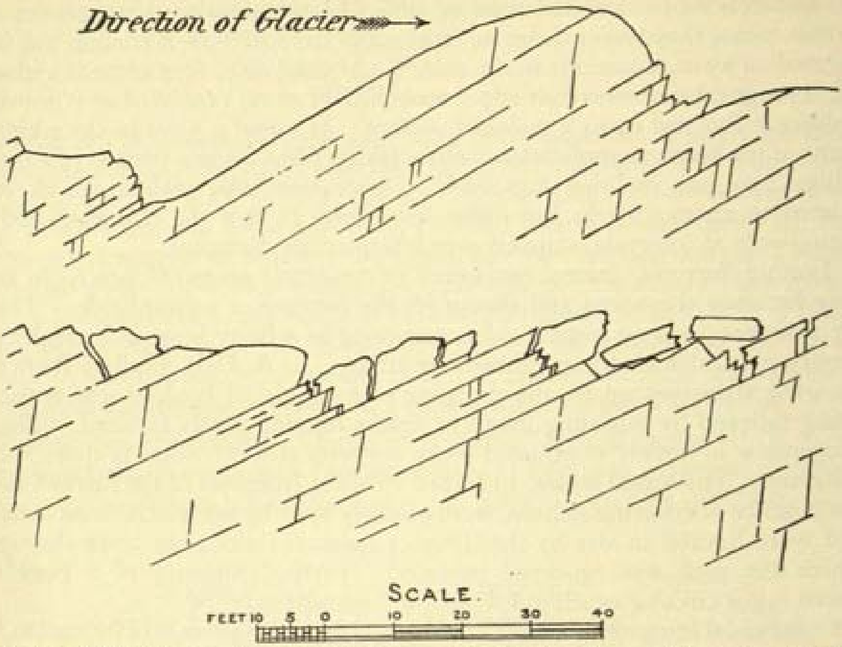


FIG. 50.—Two sections across the trench in constructing the Vyrnwy Dam, North Wales. J. A. Picton, *P. Lpl. G. S.* 6, 1889, p. 90, figs. 3 and 4.

the ice passed off or along coastal cliffs or operated on jointed rocks in bluffs, escarpments, or steep valley walls, e.g. the Palisades of New Jersey.[80] It was very effective on sedimentary strata which dipped off igneous rocks, as in the glint lakes around the Canadian Shield, in the St. Lawrence valley below Quebec, in the sounds east of Hudson Bay, and in the North Channel of Lake Huron.[81]

That plucking is a force to be reckoned with when the ice-flow is sufficiently rapid to remove the weakened material has been demonstrated on modern glaciers[82] (some may be due to frost acting on rocks exposed during a retreat[83]). Careful investigations in glaciated regions frequently show pieces broken off from the lee of roches moutonnées or other steep faces still in apposition or, as Sefström[84] noticed, dislodged to some distance; these can be referred to their exact place of extraction by accurately matching the positive and negative shapes (see p. 229). Sometimes the rock is seen to break into pieces[85]; "block tails" may result.[86]

Plucking *v.* abrasion. Early geologists, with few exceptions,[87] were either unaware of or underestimated the importance of plucking. More recently, plucking (variously styled "detraction",[88] "detersion",[89] "rough-hewing"[90] or *splitternde Erosion*[91]) has been increasingly emphasised, notably in America,[92] though a few still deem it exceptional and unimportant.[93] The great majority of those who have worked in heavily glaciated countries like Fennoscandia, west Ireland, the Alps and North America, rightly regard it as much severer than abrasion.[94] While the latter steadily diminished as its work was performed and yielded sands and muds, plucking liberated the boulders which abound in the drifts and certain big block moraines in Norway, Greenland and the Alps which can only have come from subglacial

sources,[95] and continued to act until the joints had become too tight and widely spaced in depth.[96] Its action is confirmed by the following: the rawness of the lee faces[97] where much more erosion took place than on stoss (abraded) slopes—in east Massachusetts, granite has been glacially torn off to a depth of 3–4·5 m from the stoss sides of hills and of up to 30 m from lee slopes[98]; ice does not work regardless of structural differences; rocks, relatively soft in the scale of hardness, project above surrounding harder and jointed rocks—examples are olivine rocks in hard gneissic granites in Finland[99] and the soft crystalline limestones in the silicate rocks of central Sweden[100]; hard but jointed dykes on the Cornell University Campus have been rapidly eroded[101]; in Maine, a massive crystalline schist with rectangular jointing was removed and adjacent soft shales were left standing[102]; and rock-basins have been excavated in crystalline rocks above the Malm limestone barrier of the Kirchet near Meiringen.[103] The abundant Fennoscandian lake-basins may be due to plucking[104]; others may mark the sites of plucking, their barriers being places of abrasion.[105]

Plucking operated at a rate which depended upon the thickness of the ice. Favoured by small vertical pressure—the subglacial rock-face in a tunnel beneath the Mount Collon Glacier was found to be separated from the ice by a 2–4 m gap[106]—it was confined to the margin or to the later stages when the ice was thin[107]: tensile fracture of the ice took place in the lee of roches moutonnées where the rock was not supported by pressure. At the onset of glaciation, when the ice was predominantly thin, it was facilitated by severe antecedent frost but was hampered by the excessive quantities of residual soil which tended to further abrasion. At all stages its steady supply of material facilitated abrasion downstream.[108] During maximum glaciation, when the vertical pressure was high and the ice plastic, abrasion reached its peak and the ice descended in the lee of obstacles to abrade rather than pluck[109] (see below).

Mechanism. Plucking is accomplished in many ways. The more important include (*a*) drag resulting from flow-pressure in fairly rigid ice,[110] especially at rock-projections or on ice-fronts; (*b*) movement of ice firmly frozen to the rock,[111] a method aided by "exfoliation" along vertical joints parallel with the lee-face and implied in the crescentic gouges and conchoidal fractures but dependent upon the adhesion of ice and rock and their relative strengths of cohesion—at the crest of a lee declivity, as of a roche moutonnée (see below), tension raises the melting point, and the sudden diminution of pressure results in a firming of the ice and its freezing adherence to its bed, the measure of the adherence being at least that of the tensional strength of ice, namely about seven tons per square foot; (*c*) thrusting of ice, drift or erratics along structural planes on the impact side and the prizing off of rock-pieces,[112] e.g. about Liverpool, in the Chalk of the Yorkshire Wolds, in the decomposed granite of Anglesey (see p. 364) and at Kinnekulle, Sweden; (*d*) repeated freezings and thawings at the sole in consequence of varying pressure[113] (see p. 301); and (*e*) frost acting during oscillations, as in front of the Hintereisferner.[114]

3. *Roches moutonnées*

De Saussure[115] gave the name roche moutonnée to the distinctive, rounded forms which abound in glaciated terrain (he himself failed to associate them

with ice) and give the effect of a thick fleece or the wavy wigs styled *moutonnées* in his day (they were slicked down with mutton tallow). It was again used by B. Studer[116] and became general after Agassiz[117] adopted it in 1840. *Roches bosselées* and *roches nivellées*[118] have been suggested for the large planed surfaces that resulted, it was thought, from more rapid flow.

Roches moutonnées (Ger. *Rundhöcker*), like striae, have been abundantly described and illustrated in glacial literature; they are elementary erosional forms in the glacial landscape and have occasionally been seen associated with modern ice.[119] Their gracefully moulded contours, often oval in plan, are ever varying as to dimensions and endlessly repeated as to their features. They range from low shields to steep-sided eminences and considerable hills like Holyhead in north Wales, Sheffield Pike in the Lake District and Malin Head in Co. Donegal or mountains in Norway, northern New England, e.g. Mount Monadnock or the Appalachian Plateau—they form what has been termed an *Exarationsrücken-Gelände*[120]—but in any particular area are usually of equal size. They are best developed at low levels or on flat ground and produce Penck's *Rundhöckerlandschaft*[121]; such is the vast sea of roches moutonnées on the hills and plateaux of central New York, those on the Alpine passes, and those which cover whole valley floors (Ger. *Rundhöckerflur*). Flatter and broader shapes decorate steep slopes or the sites of ice-falls. Their iceward sides are well rounded and severely scoured if the pressure crowded and compressed the structure, and were moulded by rasping or by overcoming the cohesion of the cementing material. Striae curve round a boss if it is high and narrow but arch over low and broad summits.

The lee side, which the parabolic curve of flow more or less protected from flowage pressure,[122] was usually roughened and hackled by plucking. Well-jointed rocks like granite may have been carved into steps.[123] The renewing process is visible in the opened joints or the rocks displaced horizontally or vertically through a few centimetres or a much greater distance. The moving ice pulled the joints open, stones slipped as wedges into the widened crack, and pieces were finally dragged off. The lee may be shattered or completely brecciated[124] or a whole roche moutonnée may be wrecked with the ruins of its own well-striated surface scattered about.[125] A land with pronounced joints, e.g. the granitic Åland Islands,[126] may be rugged all over.

This marked contrast, noted as early as 1799 by C. de Lasteyrie,[127] between the stoss and lee sides, terms originally used by Sefström[128] (*Stötsida*, *Läsida*)—alternatives are "strike",[129] "struck",[130] "shock",[131] "impact",[132] "onset",[133] "scour" and "pluck"[134]—explains the contrast so many glaciated valleys and hills present when viewed with or against the flow (pl. IXa, p. 256).

Related to the roche moutonnée is Chamberlin's "advance cone",[135] a half cone in bas-relief on a glaciated rock which has its apex pointing upstream, its base rough and uneven.

The roche moutonnée's shape varies with the texture and cohesion of the rock and with the nature and alignment of the structure. While it is beautifully fashioned in gneisses, granites, dolerites and diorites, a perfection erroneously ascribed to a primary tendency to concentric jointing,[136] it is poorly displayed in thin beds like flagstones or soft sandstones which are almost incapable of receiving it. Sedimentary strata dipping icewards readily adopt the form, the steep lee coinciding with normal escarpment faces or with foliation or cleavage planes. Kjerulf,[137] a persistent advocate of the insignificance of glacial erosion, ascribed the roche moutonnée's asymmetry

to such desk structure in which structure prevailed over ice-sculpture. Such protuberances are apt to be broad and squat. Structural influence is also seen in granites with sheet jointing[138] and in the reversed moutonnée forms, i.e. those with steep or "false" stoss sides.[139]

The lee side, especially in steep valleys, is frequently dressed like the impact side, the entire boss being longitudinally convex (Ger. *Vollrundhöcker*). This abnormality was connected possibly with the angle of slope[140] or the nature of the rock[141] but more probably with the severe ice-pressure[142] at the peak of glaciation when the ice, sagging behind the obstacle, abraded the lee. The plastic conditions of this time are also seen in the undercut flanks of roches moutonnées.[143] Thus crag-and-tail features are mainly restricted to low lands, and lee sides among the mountains of Skye are polished but in the submontane belt are craggy.[144] A similar change was early noticed in passing from the Scandinavian divide to the Norwegian coast[145] and in the Alps where roches moutonnées attain a far greater expression towards the heads of glacier-occupied valleys.[146]

Both ends are rounded if, as on the central Swedish iceshed, the ice flowed from opposite directions at different periods (see p. 668); the smoothness imparted to the lee hindered plucking during the reverse movement.[147]

The contrasted agencies of scour and pluck produce striking differences in escarpments. Those facing the ice, as south of Lake Erie and Lake Ontario, may be polished and grooved and have bevelled and striated edges, while those directed downstream, particularly if severely jointed, may be plucked into half-disrupted ledges with boulder-accumulations so vast that they simulate a retreat moraine.[148] The steep northern face which the dip slope of the Whin Sill presents in Northumberland may be linked in part with ice-pressure from the north.[149]

Although in the orthodox view roches moutonnées most often coincide with obdurate rocks and were predisposed by local resistance[150] or irregular preglacial weathering,[151] they are more probably not remainders nor signs of imperfect ice-erosion but like drumlins (see ch. XIX), which they resemble in their general elliptical plan and parallelism with the flow, are stable forms beneath an ice-sheet and possibly associated with glide planes.[152] They are end-forms which ice continuously reproduces, and are most prominent on transfluence passes and at the heads of valleys where the ice fashions ever new forms in accord with its dynamics, i.e. where glacial erosion was most vigorous and longest continued. Although they may have been moulded by "stationary" waves in the ice-sole,[153] similar to those which wind sometimes creates in sand,[154] or to the turbulent flow of low-viscosity fluids, such wave-flow seems irreconcilable in glaciers with the viscosity of the ice and with the trend of the striae on vertical faces. It is most unlikely that they or drumlins are merely a moulded network of subglacial stream channels, as postulated.[155]

The roche moutonnée form is attributed to the reduced erosive power on the uphill side of the rock and the thaw-freeze process on the lee side.[156]

Selective erosion. Selective erosion,[157] recognised by Lyell,[158] was at work along structures or joints or along the strike when this was parallel with the ice-flow—the elongated ridges called *vaara* in Finland were fashioned in this way.[159] It operated too on homogeneous rocks if these varied in their direction of dip[160] or were overridden by ice shod in adjacent layers with different quantities of material.[161] In bedded rocks, gneisses and slates, the ice ploughed and frequently undercut the banks.

This erosion was selective, for instance, along dykes[162] bordering the ice-fields of the Uinta Mountains or along fault-lines[163] in Finland and the Bohemian Forest and on the floor of the Baltic Sea. It initiated lakes along soft rocks,[164] as in the Laurentian Highlands of Canada, or joint planes[165] in Sweden, south-east Norway, the Åland Islands and the Rosses, Co. Donegal.

When the ice flowed parallel with jointing it gouged out any pre-existing furrows and strengthened a weak relief.[166] It smoothed and striated the surface and moulded it into long rolls, frequently on so large a scale that the country was "ribbed" or "fluted".[167] In other cases, it terraced or furrowed the valley sides along the outcrop of soft or well-jointed beds,[168] the broader benches sometimes having an undrained depression at the back. Such control may give rise to homoclinal strike valleys with rather marked asymmetry of transverse profile. In rocks and structures athwart the flow, equilibrium was quickly established and selective erosion was almost inappreciable[169]: slight bevelling marked the onset and slight plucking the lee side.

4. *Crags and Tails*

The feature of crag and tail, first described by Hall[170] from the Midland Valley of Scotland, usually centres about some hard rock,[171] such as an igneous sill, neck or dyke. The abundance of necks accounts for the very fine crags and tails to be seen in the Midland Valley of Scotland as well as for their early discovery there, e.g. Calton Hill, Castle Rock and Arthur's Seat in Edinburgh, North Berwick Law and Traprain Law in the Lothians, Largo Hill in Fife, and Necropolis Hill, Glasgow.

The impact side, which is scraped bare, is commonly steep or precipitous and, as at Castle Rock, has a horse-shoe shaped valley half encircling its base and extending leewards as lateral grooves which gradually diminish in cross-section, e.g. Princes Street Gardens and Grassmarket. The frontal groove, sometimes undercut or occupied by a rock-basin,[172] may be missing if the obstacle had a lifting effect and the ice was fairly rigid; the "precrag" accumulations of the closing stages of glaciation[173] probably arose in this way.

The tail frequently descends from the very summit of the hill in a smooth, gentle slope (pl. IXB, p. 256) whose length depends upon the height of the boss; it may range from several kilometres to only a few metres or centimetres. Grains of garnet or quartz in schist, quartz pebbles in slate, chert, flint or fossils in limestone, quartz-augen and pebbles in conglomerates and amygdules in basalt provide miniature crags and tails[174] ("ice-shadows"[175]; "knobs and tails"[176]). Island-bosses in the sea have built tapering shallows,[177] as in the lee of Gotland and of Bornholm (Rönnebank).

The tail may be solid or may consist of drift or preglacial soil preserved in the *morte-espace* of reduced ice-pressure or stagnation.[178] Examples of tails of unremoved solid rock[179] are connected with Stirling Castle Rock, Castle Rock, Edinburgh (Upper Old Red Sandstone and Carboniferous sandstones and shales under High Street), essexite and nepheline syenite at Mount Royal, Montreal, and limestone in the lee of dykes in the St. Lawrence region. Many owe their shape in the main to preglacial weathering[180] or to desk-structure, as Playfair[181] observed. Calton Hill, Edinburgh, for instance, has its dip slope to the east or lee, its escarpment to the west or impact side. An iceward dip, as in Corstorphine Hill near Edinburgh, makes the crag rudimentary.

While tails of preglacially weathered material are infrequent,[182] tails of drift are very common; for crags resemble boulders in a river bed with tails of sediment pointing downstream—the resemblance led Sir J. Hall to invoke strong currents for the Scottish crags and tails. Debris saturated the stagnant ice and ultimately displaced it. As examples may be cited Arthur's Seat, Edinburgh, Möen, Gotland and Bornholm, and the horsts of Jasmund and Rügen in the Baltic.[183] Drift, heaped up in the angle between confluent valleys, often had a similar origin.[184]

5. *Subglacial Disturbances*

Character. While solid rocks as a rule were striated and otherwise glaciated, other rocks, especially if they were soft, were often disturbed by the ice—in the case of the *Braunkohle* of Germany to a depth of 150–200 m.[185] Gravels are apt to be contorted, and solid rocks exhibit a minute jointing which passes into brecciation: in the East Anglian chalk, this is accompanied by a crumpling of the flint layers and by thrust faulting. Clays are squeezed into folds, and sands are driven into the arches. The folds become steep or vertical and finally break into fragments which ride forward. They sometimes lie on a particular horizon between beds which remain horizontal.[186] Over an area measuring 12 km by 8 km in west Fläming, the Oligocene and drifts form slightly curved and parallel anticlines and synclines up to 30 in number, bifurcating and re-uniting with a distance from saddle to saddle of *c.* 100–300 m.[187] Elsewhere in north Germany there are folds more than 1 km long. The Saale ice-sheet produced in Emsland a schuppen structure[188] in the earlier drifts and underlying Oligocene and Eocene beds to a depth of 150 m, which was the depth of the permafrost; the folds and schuppen axes are shown in the text-figure[189] (fig. 51).

Distribution. Disturbances of this kind, which have long been known,[190] are widespread in the Mesozoic and Tertiary strata of eastern England,[191] as in Lias clay at Blisworth, Great Oolite near Towcaster, Oxford clay at Huntington, and Chalk near Flamborough Head and Hitcham; in north Germany[192] in lignite, Muschelkalk, Chalk and Tertiary, especially in the area of the Oder Lake; in Scania and south Sweden, as described by E. Erdmann, O. Gumaelius, A. G. Nathorst, D. Hummel and O. Torell; in Denmark (see below), Holland,[193] and east of the Baltic, as near Leningrad.[194] They have been occasionally discovered in Switzerland,[195] Iceland[196] and Russia (see p. 259) and in several North American localities,[197] including Alberta (Cretaceous rocks), Martha's Vineyard (sharp anticlinal folding in Miocene clays) and Long Island, whose flexures are 50 miles (80 km) long, several miles wide and 50–100 ft (15–30 m) high.

Glaciostatic origin. While some may doubtfully and locally result from slips,[198] a falling water-table,[199] an epiglacial rise of the isogeotherms,[200] local ice dropping blocks or grounding in extraglacial lakes,[201] decalcification, solifluxion (see p. 567), or melting of dead ice (see ch. XXI), widespread disturbances in solid rock cannot be so explained. Glaciostatic or tectonic forces only can account for them.

Advancing ice may be likened to an overthrust sheet and its ground-moraine to friction breccia or mylonite.[202] It striated and polished the rocks and plucked and shattered them to a depth of several metres, displacing them by small faults.[203] It inserted tongues of boulder-clay or boulders along

their structural planes (see p. 364), e.g. the granite blocks half buried in the Devonian clays of Dorpat.[204] It stripped off patches of the preglacial rubble and incorporated them in the drift, together with rolled up wisps or pillow-shaped masses of the subjacent clays, and kneaded together the glacial and preglacial material.[205] It generated folds, overfolds, isoclinal folds, and thrusts perpendicular to its flow[206]; brecciated, slickensided, crushed and

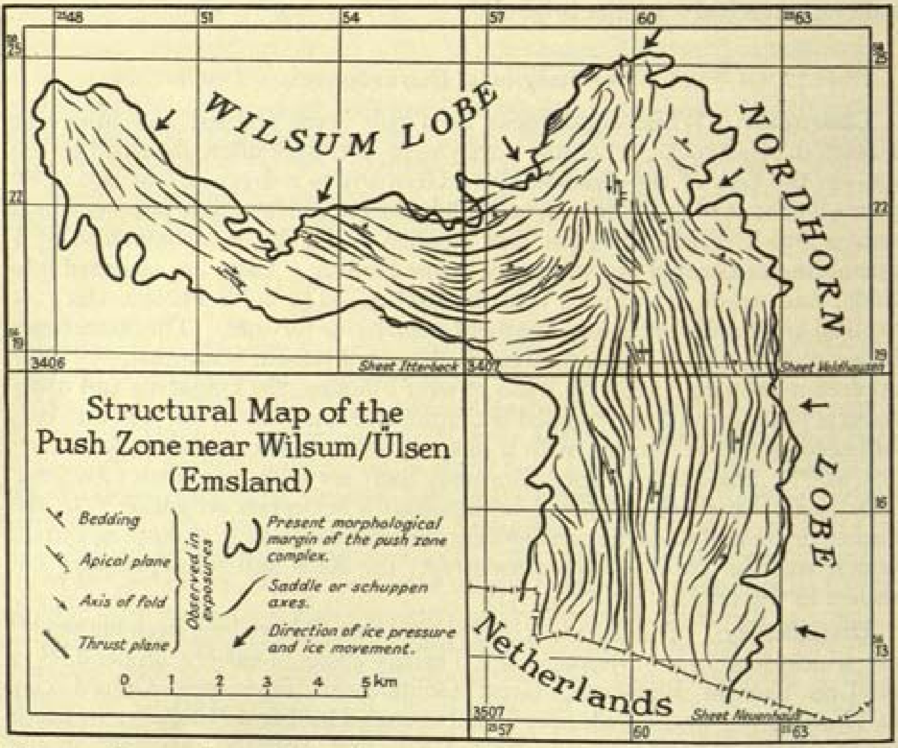
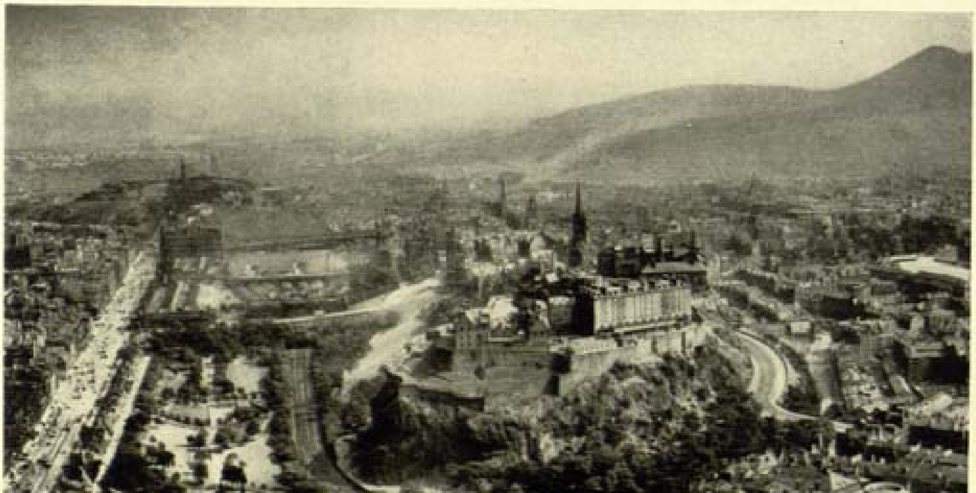


FIG. 51.—Structural axes of the schuppenzone in Emsland, west Germany. W. Richter *et al.*, *Z. D. G. G.* 102, 1950, p. 70, fig. 4.

shattered the rock; imparted a *schuppen* structure to the drift[207] and the underlying beds, as near Hamburg,[208] to the browncoal of Germany[209] and to the Chalk near Chesham (Buckinghamshire) which it tore into long strips and piled up one upon another[210]; and displaced large blocks composed of several layers, arranged either in normal succession or folded and inverted[211] —a bore near Rostock passed through six *Schollen* intercalated with drift.[212] These masses by their ruin in further transport may explain the local abundance of certain fossils or erratics of a particular or rare kind of rock[213]—three boulders of Shap granite at Balby, Adwick and Royston in south Yorkshire, which have a vein of felspar running through each of them, were formerly one boulder.[214] An inverted mass of Gault and Greensand, *c.* 230 m long, at Leighton Buzzard[215] was transported *en bloc* from the north and overturned when its sands and clays were frozen hard.[216] Alternatively, its separate units were piled up in inverse order one above another.[217]

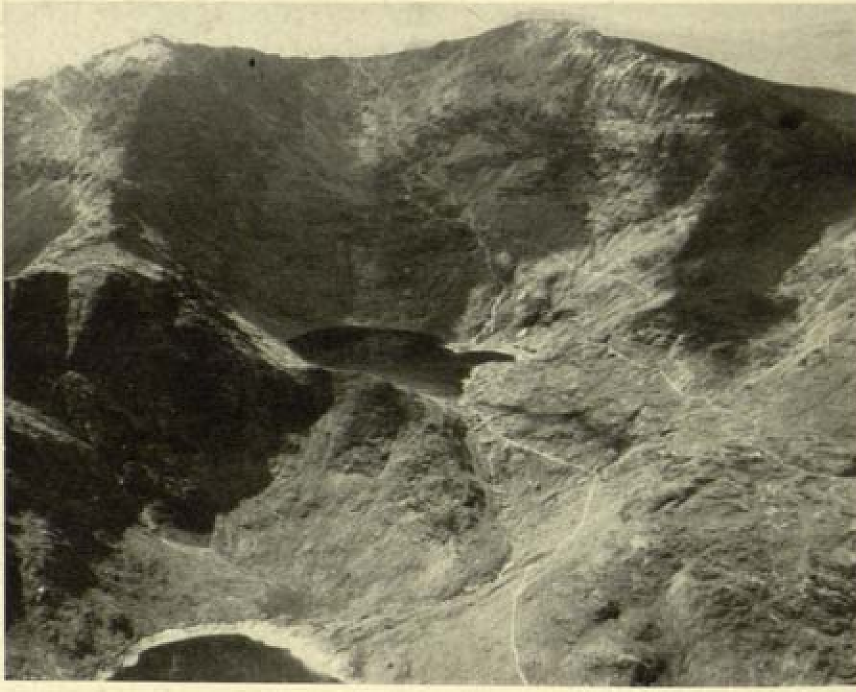
These disturbances were apt to occur if the ice-flow was impeded by irregular ground, as on the faulted country of Holland,[218] by ascending

A. View looking from south side of Mt. Washington toward Mt. Monroe (5,385 ft.), White Mountains, New Hampshire, shaped by ice overriding from north-west (*right*) to south-east (*left*) [R. J. Lougee]



B. Edinburgh Castle Rock showing upstream crag encircled by horse-collar groove and tail to the east [*The Times*, London]

(*facing p. 256*)



A. Cwms of Snowdon looking westward over Llyn Llydau and Glaslyn
[J. K. St. Joseph : Crown copyright]

B. Cirques about Striding Edge, Helvellyn, Lake District
[J. K. St. Joseph : Crown copyright]

(facing p. 257)

slopes,[219] as south of the Baltic depression or north of the Trebnitz Gebirge[220] (where faulting has also been invoked) or along the sides of valleys, e.g. near Posen, in north Germany and in East Anglia (see p. 992).

Glaciostatic disturbances may be attributed to push effects at the edge of the ice during a general advance[221] or, more probably, during a general retreat[222] when the ground had thawed and the folds, etc., were left undisturbed. It has been concluded that the permafrost was thawed when the ice advanced on to it and so was eroded[223] and that lines of blocks in the lower part of the till represent the plane between frozen mass above and unfrozen mass below.[224] Studies about Breslau[225] suggested that beds were folded when the ground was unfrozen and the ice was thin but were thrust and planed off if the ground froze and the ice was fairly thick. Large-scale disturbances, however, were probably submarginal and not formed under the thick ice of maximum glaciation.[226] They have been ascribed to torsion, e.g. between the east Baltic and Småland ice.[227]

These crumplings, foldings and displacements are well displayed in many places, including the New England islands of North America.[228] The finest are on the south coast of the Baltic, as at Rügen and Möen and in the Danish "Klints" (e.g. Möens Klint, Lönstrup, in north Jutland, Ristinge on the west side of Langeland, Rögen on Fyn), and have occasioned much controversy. Ice-pressure was suggested by F. Johnstrup[229] two years before O. Torell's paper advocating the glacial theory for north Germany (see p. 627). It was later advocated by many others[230] who emphasised that the planes of dislocation were irregular and restricted to the upper 20–40 m and dipped towards the direction whence the ice came; that the included masses, such as chalk, in the drift tended to form gigantic "eyes" and tail off at either end into streaks; and that like structures occurred in Schleswig and all parts of the North German Plain[231]—drift underlies 200 m of chalk in Pomerania.[232] Gripp[233] interpreted them as stau-moraines when Jasmund and Möen were nunataks and E. Philippi,[234] in a documented review, regarded them as glacial thrusts falling northwards. Spitsbergen provides modern parallels.[235]

G. Slater[236] maintained that the movement at Möens Klint was subglacial and differential as proved by the strips of drift intercalated in the chalk and the latter's internal structure, and by the typical anticlines which are distorted into flow-overfolds parted by thrust planes (see p. 398; fig. 52).

Tectonic origin. While this glaciostatic explanation is the more likely one, German geologists have on the whole favoured the purely tectonic or non-glacial origin,[237] first expressed by Lyell[238] and accepted during the present century by an increasing number of Danish geologists[239] for Möens Klint, Lönstrup Klint, Ristinge Klint, Rögen Klint, Vendsyssel and Scania—the earlier view favoured plutonic forces.[240] Thus the disturbances are parallel and coincide with areas of faulting in earlier periods[241] or to-day[242] (slips ?), and the drift and chalk meet along smooth and regular planes. The date of the faulting is disputed: preglacial,[243] postglacial[244] or interglacial[245] are the dates assigned. The faults have also been thought to have taken place during the advance of the ice and to have recurred with each glaciation from tangential pressure exerted by the ice upon an unstable terrain[246] and from an awakening of inherited tectonic tendencies.[247] Interglacial faulting has found most favour[248] ("Baltic faults"[249]; "Baltic dislocation phase"[250]); it accounted for certain characteristics of the upper drift, namely its high lime

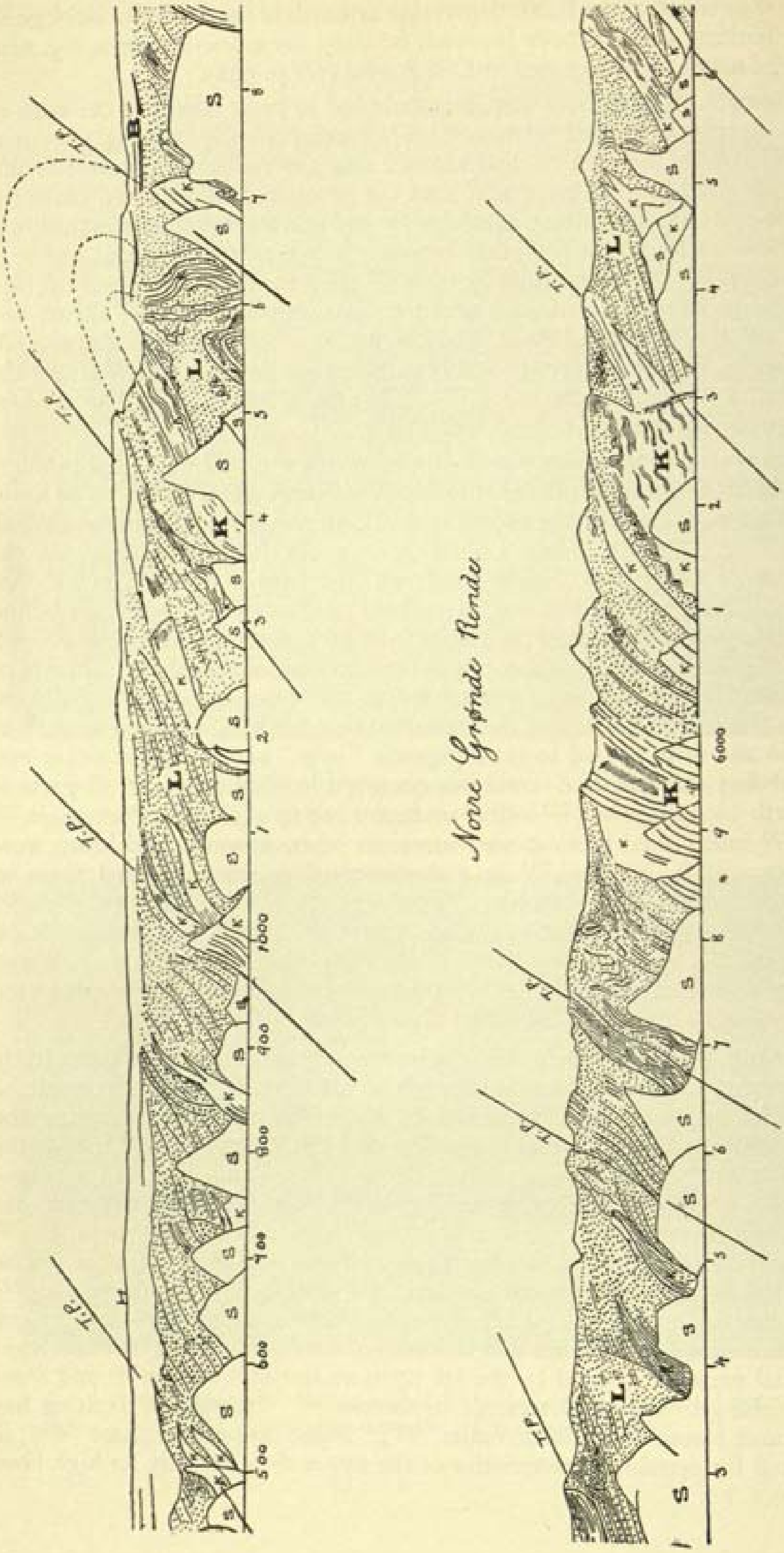


FIG. 52.—The disturbed glacial deposits near Lönstrup, Denmark. G. Slater, *T. R. S. E.* 55, 1928, pl. I (part of).

content in Schleswig-Holstein, its undisturbed passage over the dislocated lower drift, and the distribution and great size of the erratics in north Germany (see p. 364). A middle view invoked both glacial and tectonic action.[251]

Russian geologists are similarly divided upon the cause of the disturbances that are known from various parts of European Russia,[252] as near the Valdai Hills, near Lake Peipus and in the Volga and Dnieper country. In the region of the Valdai Hills Carboniferous masses have been pushed over glacial deposits for a distance of *c.* 100 km and smaller masses occur in the Smolensk-Moscow ridge north-west of Moscow. West of the Dnieper, they occur over a region 70 km long and 35 km broad and include besides Quaternary deposits, rocks of Jurassic, Cretaceous and Tertiary age. In favour of the glacial origin is the small altitude of the zone compared with its great width.

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Chapter XII

ROCK-BASINS

A reality. It is now agreed that closed rock-basins, other than solution lakes or tectonic lakes like those of East Africa or the Dead Sea, do exist. The occasional denials[1] are unjustified; for the hypothesis of detrital dams has succumbed to careful investigation. Nevertheless, some basins, as in Snowdonia,[2] are not demonstrably rock-bound and others whose outlets are over rock-rims, as in the English Lake District,[3] may have preglacial outlets concealed by drift. Most lakes had a composite origin; their lower layers alone are rock-bound, the upper ones being ponded by moraines or drift which crowns rock-ridges and both heightens the lake's surface and determines its configuration and extent. If their dams were removed, Lago di Como would lose *c.* 150 m of its depth, and Lake Constance one-third of its area.[4]

Shape and size. Shape and size, together with local structure, are important clues to the origin of rock-basins—W. Halbfass[5] has catalogued the dimensions of the world's main lakes. They vary in plan from the oval cirque-lake to the elongate *Rinnensee* (see p. 241). Their sides are steep and lack sinuosities and their isobaths portray the downward continuation of the subaerial slopes. The bottom is more or less flat but may be divided into subsidiary basins by subaqueous barriers which may be moraines,[6] e.g. in Lake Lucerne, Zürich See, Lago di Lugano and Lake Lomond. It is often spoon or shovel shaped, the valleys to use W. D. Johnson's apt phrase being "down at the heel".[7]

Most lakes were originally deeper than now since they are floored with alluvium of an unknown thickness which has sometimes made the central parts remarkably even[8] and has only occasionally been penetrated[9]; it is about one-quarter of the original depth of the Freibergsee in Allgäu[10] and may be 80 m in Lake Geneva.[11] Its unsuspected great depth led to the disaster during the cutting of the Lötschberg tunnel.[12]

Generally speaking, the smaller the lake, the greater proportionately is its depth.[13] Thus cirque-lakes are relatively deep[14]; this has been demonstrated in Sutherland, Snowdonia, Co. Wicklow, the Lofotens, the Alps and the Pyrenees. Their quotient of depth to length exceeds that of other lakes; it is as high as 1:8·58 in the Hohe Tatra.[15] In "cryptodepressions"[16] the floor is below sea-level: it is necessary, for example, to go 400 km from the mouth of the River Po to south of Lissa in the Adriatic before the depth of Lago di Garda (= 346 m) is equalled and 120 miles (*c.* 200 km) into the Atlantic Ocean to attain the equivalent depth of Loch Morar (987 ft: 301 m).

Much misconception would be removed if, as Ramsay[17] urged, lake-profiles were drawn to natural scale. The water in the vast majority of cases is a mere film. Even the deep subalpine lakes, as tables and sections show,[18] are insignificantly thin compared with their length (see p. 276). This is true too of the Scottish lakes[19]; thus Loch Lomond is 176 times longer than its greatest depth and Loch Ness 136 times, and the ratio of length to depth in Loch Morar, the deepest of the British lakes, is 61:1. . Exaggerated profiles

are largely responsible for the statement that ice does not "dig holes"[20] or for the view that lakes are pot-holes or erosion basins drilled by subglacial streams (see p. 240).

It is, therefore, not the depth but the gradient of the reversed slope (Fr. *contre-pente*) that is vital. This is 1:300 in Ammersee and 1:190 in Starnberger See[21] and is only 2° 21′ in Lago Maggiore,[22] the second deepest of the subalpine lakes. It varies in Swiss lakes between 19‰ in Lake Geneva and Zuger See to 53‰ in Lake Lucerne.[23] It is 1° in Chiemsee, which is typical of other Bavarian lakes,[24] and ranges between 7‰ and 23‰ in Scandinavia.[25]

Origin. Early Alpine monographs, such as those of J. Payer on the Ortler Alps, K. A. v. Sonklar on the Stubai Alps, and A. Waltenberger on the Rhätikon, Wetterstein and other mountains, were purely descriptive of the orography and orometry. Any explanations they offered reflected the catastrophic views of A. v. Humboldt, L. v. Buch, É. de Beaumont and B. Studer, the teachings of J. Hutton and J. Playfair being either unknown or disregarded. Once the origin of the features came to be discussed (O. Peschel[26] was one of the first to do this) widely differing opinions arose. Hutton had only been able to suggest three explanations: landslips, earth-movements and solution.

Rock-basins, as we now know, may be hollowed out in many ways; by wind as shallow deflation basins,[27] e.g. in Pomerania, central Asia, New Zealand, the North American prairies and on high granite plateaux in the Scottish Highlands; by unequal weathering under peat,[28] as in the English Lake District; by solution in limestones,[29] as affirmed for the tarns of the Canton Ticino (cf. p. 266), some Irish lakes and the Limestone Alps including the Triassic area of the Adamello group; by subsidence,[30] as in Albania; by swirling waters[31]; by anchor-ice[32]; or by ice expanding in crevices and lifting out the loosened blocks.[33] Nevertheless, the vast majority were not made in any of these ways: the forces are quantitatively inadequate. There, however, agreement ends for views diverge widely. Hence the problem of the rock-basin.

It is, of course, a truism that the genesis of each lake must be sought separately with respect to the shape of its floor and shore and its relation to local relief, lithology and structure. Yet the question resolves itself in practice into deciding between two forces, tectonic movement or ice.

Tectonic. It is widely claimed that rock-basins were made by the first of these, i.e. by folding, torsion, warping or dislocation. Earlier geologists[34] were generally of this opinion and their successors have invoked flexures across valleys,[35] as for Lake Geneva and the lakes of Salzburg, north Italy, Dauphiné, the foot of the Jura Mountains, and the Alps in general,[36] as well as for some Scottish,[37] Norwegian[38] and Pyrenean[39] lakes. But the basins are frequently composite and, as in the case of Lake Geneva,[40] without detectable flexures; those in adjacent valleys or even in different parts of the same basin require different causes and movements bewilderingly complex. Furthermore, the movements must have been comparatively recent and so rapid that erosion could not keep pace with them. A hypothesis involving such a medley of unlikely assumptions stands self-condemned.

Faults and jointing have been deemed responsible for Lake Ladoga,[41] König See[42] and Loch Morar[43]; for the lakes of Sweden[44] and Finland[45]; and for some in Bavaria,[46] Italy[47] and the Vosges.[48] Dislocations and

tectonic movements are suggested too for the Jura lakes[49] and for certain subalpine,[50] Bavarian[51] and Pyrenean[52] lakes. Recently, they have been repeatedly demanded for Lake Constance[53] after M. Schmidle[54] had shown that the lake lay between longitudinal faults and had horsts rising from its floor. The faults were subsequent to the Mindel glaciation and are still active as recent earthquakes[55] and fine levellings[56] prove.

Nevertheless, we must recognise that while dislocations, joints, dykes and junctions of hard and soft beds have largely controlled the direction of lakes and their branches, these lines of weakness have merely guided erosion[57] as in Bavaria, Fennoscandia and the Scottish Lewisian country. Lakes formed by selective glacial erosion have been termed "roxen lakes",[58] after Lake Roxen in south Sweden, or glacial lakes of the Finland type.[59]

Preservation hypothesis. Lakes, it has been supposed, mark the sites of preglacial or Tertiary depressions kept free from sediment by Pleistocene glaciers. This "preservation hypothesis" has often been applied to the subalpine lakes,[60] especially in north Italy, to some Bavarian lakes,[61] and to the glint lakes of Swedish Lapland.[62] It was widely invoked by E. Reclus[63] and revived in a modified form for Switzerland by Heim[64] (see p. 279) and by R. Staub[65] who postulated masses of dead ice in the *Randseen* and other lakes, on the evidence of backward inclinations of lateral moraines and deltas in the upper parts of the lakes and of an early glacio-lacustrine clay on the lake-floors, the ice-masses persisting for periods depending upon the local conditions, e.g. precipitation, mist and winds (including the foehn). Yet the hypothesis fails to explain the fluvioglacial material above as well as below the lakes; their restriction to glaciated terrain; and the absence of preglacial alluvium under the drift.[66] Its main weakness is its failure to solve the origin of the lake-hollows: it relegates these unexplained to preglacial time.

Ice-erosion. The general acceptance of the glacial theory for the phenomena of the drift, particularly in Switzerland, made possible a further advance in the study of glacial forms. Ramsay[67] in Europe and Logan[68] in Canada almost simultaneously conceived the idea that ice had scooped out lake-basins. Attracted by its simplicity, a few British scientists[69] immediately welcomed the theory. But the balance of opinion was hostile and prominent geologists of the day, led by Lyell[70] in England and by Heim on the continent of Europe, rejected it. The theory passed through an eclipse which was only ended by its renewed championing[71] by Penck and Wallace. The steady support of the Scottish school in Britain,[72] of Davis and Gilbert in North America,[73] and of Penck and Brückner on the mainland of Europe[74] led to its revival and indeed to its extension at the end of the century to the valleys of which the basins are an integral part (see ch. XIV).

The theory has been held for most lake-regions, e.g. the English Lake District,[75] the "finger lakes" or "piedmont lakes" of Sweden,[76] the Alps,[77] Pyrenees,[78] Black Forest,[79] High Sierras of California,[80] New Zealand,[81] Canada's glint lakes[82] (eroded along the boundary between the Archaean and Palaeozoic strata), and for limestone regions[83] like the Canton Ticino and central Ireland.

Proof. Glacial erosion of preglacially weathered material and of some solid rock is in the vast majority of cases the most satisfactory theory. It harmonises most readily the greatest body of observed facts. Nevertheless, direct and convincing proof is by no means easy to gather. Some of the

evidence cited is unsound: it merely establishes glacial occupation. The glaciated sides, lips and floors of the basins, revealed by draining some lakes in the Alps and Pyrenees for power undertakings,[84] strictly demonstrate only that ice occupied the hollows. Similarly, an appeal to their recency[85] logically proves a Quaternary age and not necessarily a glacial origin.

Ramsay himself, like so many geologists after him, arrived at this view by a process of exclusion. Rock-basins, although admittedly predisposed topographically, tectonically or lithologically, were not formed by synclinal folds, lines of fracture, subsidence or solution.

The most potent argument is the very striking and significant fact that rock-basins concur in their distribution with past glaciation,[86] whether this be in folded mountains like the Alps or the Karakorams, in elevated old masses like Scotland or Scandinavia, or in the basalts of the Scabland area of North America. This has been repeatedly shown, e.g. in Bavaria,[87] the Auvergne[88] and Pyrenees,[89] Bohemian Forest,[90] Greenland,[91] South America[92] and the Kosciusko region of Australia.[93] They are also restricted to those parts of valleys which the ice covered and modified from a V- to a U-shape[94] (see p. 322), as in north Italy, the Black Forest, Russian Altai and New Zealand. Lakes in the Yukon district of Alaska are exclusively confined to glaciated terrain[95] and in the Andes of Peru and Ecuador to between the present snow-line and the limit of the Pleistocene glaciers[96]: south of 41° and especially of 46° S. they occupy a series of valley or piedmont basins which in some cases, e.g. Argentino, Buenos Aires, Pueyrredòn and Viedma, lie within morainic amphitheatres of the last glaciation. The lake-zones in Patagonia converge northwards as the glaciated region narrows.[97] In North America, the southern boundary of the lake-country follows roughly the edge of the Wisconsin glaciation, as in Massachusetts, New York, Wisconsin and in Minnesota which has over 10,000 lakes. Peats marking the sites of ancient lakes are similarly related to glaciation. This is convincingly brought out in peat maps of central Europe[98] which reveal two main peat areas, the one in the peribaltic lake-zone, the other in the Swiss Plain from Lake Geneva to Lake Constance, thence over Württemberg to south Bavaria and Bohemia. In Switzerland,[99] the extent of the last (Würm) glaciation can be traced by the peat mosses. North America[100] shows a similar intimate relationship.

This coincidence of lakes with glaciation has rare exceptions. There are on the one hand the relatively few lakes which lie outside the glacial limits, as in such karstlands as the Balkans and Kentucky; in volcanic districts like the Eifel, Auvergne and south Italy; in steppe lands such as Hungary and the Great Basin region of North America; in coastal marshes; and in tectonic areas, e.g. Lake Baikal (1741 m deep) and east Africa (Lake Tanganyika, 1435 m deep). On the other hand, lakes are wanting in central areas of non-erosion, as in Lapland[101] (see p. 217), and in the more southerly situated glaciated Caucasus,[102] south Carpathians, Himalayas[103] (they are not quite lacking here) and in the southern Rocky Mountains and White Mountains of North America.

Their absence from the Caucasus is attributed to severe erosion by post-glacial rivers[104] facilitated by higher gradients and the poverty of the vegetation; to the shortness and weakness of the glaciers[105]; or to later sedimentation since the Caucasian valleys bear signs of numerous lakes up to 6 km long.[106] Their rarity in the Pyrenees, which is similarly ascribed to the shortness of the glaciers[107] or to the original shallowness of the lakes which did not survive

rapid erosion and infilling,[108] is apparently not real, for the Pyrenees have over 1000 lakes of which 942 are cirque-lakes and 99 valley lakes.[109] Their lack in the south Carpathians is likewise due to the shortness of the glaciers[110] or to the great fluvioglacial accumulations[111]; in the eastern Alps to early crustal stability[112]; in the Apennines to the occurrence of limestones[113]; and in the Himalayas and east Tienshan to rapidly eroding rivers which have prevented lakes from forming[114] or, alternatively, have filled them in[115] or—on the hypothesis that the subalpine lakes are due to subsidence (see below)—to the absence of such tectonic movements in the Himalayas.[116]

The coincidence also holds in a vertical sense. From broad, basin-shaped hollows on the plains or at the foot of mountains we pass by long, narrow lakes in the valleys to round *Hochseen* in the cirques and small hollows on the passes, e.g. the Simplon and Great Bernard passes. The lake zone rises with the glacial snowline into lower latitudes[117] (see p. 296).

The coincidence goes even further; for the lakes are seemingly proportionate in size and number to glacial severity. They decrease in the Alps[118] into the less glaciated east and diminish northwards with the glaciation in the south Alps of New Zealand. They also shoal and approach the mountains as the glaciation contracted in the German Alps[119] and in the Andes[120] where there are *Randseen* in central Patagonia and south Chile and small lakes only in the moutains at 4° 20′ N. Lat.

Additional proof is afforded by the parallelism of the rock-basins with the ice-flow,[121] as Sefström[122] noticed, even where the ice swung round or radiated fanwise—valleys transverse to the flow have few or none[123] (see p. 323); and by their situation in positions of severest glaciation.[124] Their depths, generally speaking, correspond to the size and importance of their valleys and of the occupying glaciers.[125] They are frequently, as in the Lake District,[126] below the junction of two valleys where the smaller cross-section augmented the ice-pressure. Rock-basins, indeed, often occur where flow was quickened, as in diffluence passes (see p. 332)—the Hudson River valley, in a massive highland 1400–1600 ft (425–485 m) high, has a rock-basin 765–950 ft (233–290 m) below sea-level[127]—at the confluence of glaciers[128] ("confluence basins"), and in valley constrictions[129] (due to superposition, antecedence or local rejuvenation as is seen, for example, in the unglaciated parts of the Var valley in the Maritime Alps). This analogy with the scour and fill of streams[130] accords with the law of the adjustment of cross-sections[131] and with the swelling of the ice in such positions as was early recognised from the direction of striae on valley walls.[132] Enhanced pressure and velocity augmented the scour[133] which was proportional to the velocity, either the third power[134] or directly as experiment suggests,[135] or to the square[136] or cube[137] of the depth.

According to other geologists,[138] basins occur in the widest parts where lateral friction was least, basal friction was highest, and the motion was least impeded. Complementary rock-barriers rise in narrows where spurs projecting from the sides sustained the maximum pressure.

Rocky islets may be linked with more resistant rocks (they are missing from basins of uniform rock like Starnberger See and Ammersee[139]) or with central positions where they escaped the influence of the dirty lateral ice. They may also be due to initial irregularities which rose above the debris-laden base and became accentuated,[140] or to initial shallows in the glacier-bed which, by reducing the glacier's depth, tended to become higher and higher.[141]

The basin is often deepest at its head[142]: this had the great thickness and pressure of ice (a glacier, unlike a river, comes into existence almost full grown), the longest glaciation, the most angular material, and lacked the glacial accumulations that often abound in lower stretches. Here too rotational slipping was likely to occur (see p. 118)—it also took place above the snout and below ice-falls and steps.

The glacial theory was supported for the Scottish lakes[143] which were eroded during the valley glacier phase; it was in consonance with the geological structure and glacial phenomena, with the U-shaped cross-section, with the frequent coincidence of the deepest with the narrowest parts (rock-basins are rare in the open and comparatively shallow valleys of the Moṇadhliath and Cairngorm Mountains and eastern Grampians). The lakes have separate basins and often slope most steeply at the concave bends where differential erosion was presumably most powerful.

It is corroborated by the basins uncovered on the retreat of existing glaciers[144] (French Alps, *plan*; German Alps, *Boden*; Swiss Alps, *Platten*) like the Gepatschferner, Obersulzbachferner, Plessnitzkees, and in the Krimmler Tal, Hohe Tauern, and by the descent of the floor of the Crillon Glacier, Alaska, to *c*. 50 m[145] and the Fedchenko Glacier to *c*. 150 m below the snout—this has been attributed to static buoyancy of the tongue behind the glacier outwash[146]—by the discovery, through seismic means, of a rock-basin below the Morteratsch Glacier,[147] the Concordiaplatz of the Aletsch Glacier (P. Veyret, 1954) and the Saskatchewan Glacier, Alberta (M. F. Meier *et al*., 1954), and of an occasional rock-basin between the Hofsjökull and Langjökull in Iceland.[148]

It is also confirmed by the rock-barriers (Ger. *Riegel*; Fr. *verrou*) whose convex, smooth and striated surfaces sink into depressions above and below. Plucking has often made their downstream faces steep and rugged while upstream they are mammilated and excessively scoured. They are usually pierced by a gorge, like the celebrated ones on the Aare above Meiringen,[149] sawn through by a subglacial or postglacial stream (see p. 240).

Like rock-basins, the barriers are relatively small compared with the ice-thickness[150] and are found in various glaciated regions,[151] e.g. the Pyrenees and the Antarctic and in not a few Alpine valleys, e.g. in front of the Upper Grindelwald Glacier, at St. Moritz in the Rhône and at Kirchet in the Aare valley. Many modern Alpine glaciers end just above barriers of this kind and have filled the intervening hollows with fluvioglacial debris. Others, like the Aletsch and Gorner glaciers, terminate just at the barrier, their oscillations merely swelling or lowering the ice without altering it horizontally[152]—hence the tendency of such barriers to be crowned with moraines.

Elevations of this sort crevassed the ice and may have reversed its slope; the "dimples" on some Antarctic glaciers probably mark the sites of basins behind rock-barriers.[153]

Barriers have often been discussed since Rütimeyer and Heim first drew attention to them. There can be no doubt that hard rock, e.g. limestone, or infrequent jointing favour their persistence.[154] The barrier at Kirchet occurs on an anticlinal fold[155] while many at the lower ends of rock-basins in the Scottish Highlands consist of hard rock[156] like the Ben Ledi and Ben Leny grits (Grampians), Lewisian gneiss (Ross-shire), or Moine schist invaded by a plexus of granite intrusions (Loch Shin). The persistence of limestone as barriers and the resistance of schotter to glacial erosion have been correlated

with the infiltration of waters which in this way were unable to help in abrasion as they did over impermeable rocks.[157]

Brückner[158] recognised five kinds of barrier and barrier step due to (*a*) undercutting by a main glacier at the entrance of hanging valleys and lessened erosion at the mouths of tributaries whose ice was dammed up; (*b*) selective erosion; (*c*) confluence of two or more glaciers; (*d*) varying erosion brought about by changes in a valley's width or gradient; and (*e*) prolonged halts of the ice. As de Martonne[159] has suggested, they were left where glacial erosion was reduced by the accident of preglacial relief, i.e. ruptures of slope, constrictions, emergent hard rock, elbows of capture, or lobes of incised meanders.

Barriers pass through transitions into bosses (Ger. *Riegelberg*). These rocky knobs, more or less isolated, rise where spurs have been only partially consumed (see p. 316). J. Brunhes[160] distinguished three types: (*a*) bosses at the snout, either buried and marked by a swelling of the glacier or appearing at its surface and cleaving it—instances are connected with the Upper Grindelwald and Upper Aletsch glaciers; (*b*) *Platten*, as in front of the Aletsch and Fiesch glaciers; (*c*) *Inselberge*, named from Inselberg near Innsbruck—others (better termed *Riegelberg*[161]) are found near Salzburg and at Sion in the Rhône valley.

Barriers, though regarded as entirely of preglacial development,[162] were not made by rivers which tend to efface them (see p. 506). Nor are their abundance, position or direction reconcilable with tectonic movements. They strongly argue for the glacial erosion of their complementary forms, the rock-basins. The latter often lie at the base of steps,[163] in positions analogous with "swirl-colks" at the foot of falls in rivers[164]—the basin below Niagara Falls is 50 m deep.[165] De Martonne[166] worked out a formula, not immune from criticism[167] (see p. 329), linking the friction of the ice on its bed with velocity and pressure, the latter being proportional to the cross-section of the bed and cosine of the gradient. Erosion decreases above a certain angle as the slope increases and is severest above and below breaks in the profile and in narrows. These conclusions find confirmation in the Vosges[168] where modest altitudes, relatively horizontal structures and moderate glaciation are especially favourable. They were supported by Collet[169] and by de Martonne's own observations on ground which retreating Alpine glaciers laid bare, striae and other marks being more pronounced on flats than on steep declivities.[170] Similarly, the Cascade Glacier of Alaska, protruding over the lip of its hanging valley, has been unable to remove the striations the trunk glacier made on the wall.[171]

Rock-basins, like drumlins and roches moutonnées (see p. 253), have been related to a rhythmic movement in the ice,[172] e.g. to a rhythmic alternation of plunging and rising related to the incoming of tributary ice-streams.[173]

Zungenbecken. The biggest rock-basin is the terminal basin (*Zungenbecken*[174]). It lies in the "central depression" and with the moraine (B. Gastaldi's "moraine amphitheatre"[175]) and outwash constitutes Penck's "glacial series"[176] (see p. 441). The critical line of the rock-rim may coincide with the sudden rise in the floor.[177]

These lakes, which on plains are usually round and broad and in valleys long, narrow and deep, may, like the Bavarian Chiemsee, Tegernsee and Würmsee, persist as open basins, form turf flats, or be filled with outwash, particularly if copious streams enter them, as in the Rosenheim basin.[178]

Penck[179] divided terminal basins into a main basin (*Stammbecken*), of which he recognised two types (fig. 53), and a number of branch basins (*Zweigbecken*) radiating from this, e.g. lakes Überlingen and Zell (with the valleys of Schussen and Argen) and Lake Constance. They arose where glaciers were unable to fan out upon the foreland, as in the Inn and Salzach glaciers. The rise separating the main from the branch basins may be a preglacial watershed, an elevation on hard rock, or a group of drumlins.[180]

The central basins (Fr. *bassins terminaux*) were apparently excavated just above snouts that long remained stationary.[181] On the theory that erosion is severest where the ice is thickest,[182] the diminished ice-thickness and slackened flow entailed a reversal of slope,[183] especially if the ice branched or fanned out[184] and had thrust planes in this position.[185] An additional cause may have been the increased plucking which resulted from reduced thickness[186] (see p. 251) or the exclusion of sediment by the glacier, as suggested for the central Swedish lakes.[187] Thus a morainic girdle (A. M. Hansen's "*Indsjö* stage"[188]) encircles the Scandinavian lakes[189]; the second bow of the

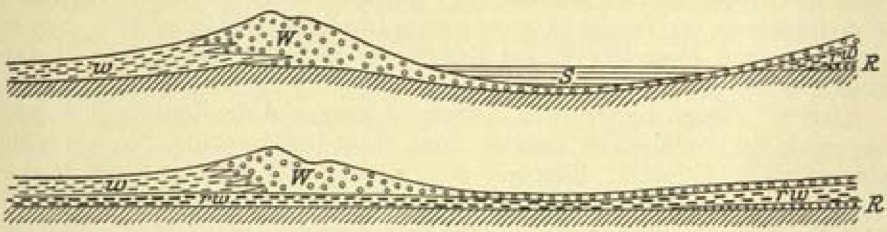

FIG. 53.—Two types of terminal basin (*Zungenbecken*). Type 1 (top diagram), Ammersee. The fluvioglacial Würm schotter (w) passes downwards from Würm end-moraine (W), which surrounds a lake basin (S) and above which lie interglacial schotter (rw) resting on Riss moraines (R). Type 2 (lower diagram), Glattal. Fluvioglacial Würm schotter (w) passes downwards from Würm end-moraine (W) and lies, like the end-moraine, on interglacial schotter (rw), which in turn rests on Riss moraines. A. Penck, *SB. preuss. Ak.* 19, 1922, p. 243, figs. 8 and 9.

Second Salpausselkä follows the southern and south-eastern end of a big lake-depression[190]; and many lakes mark the limit of the last glaciation in north-west Europe, e.g. Lake Ladoga, Lake Onega, and the Gulf of Riga,[191] for which a glacial erosive origin had early been suggested.[192] The *Zweigbecken* of the Alpine glaciation have been correlated with the Würm epoch.[193] Linstow's Diluvial Depression of north Germany (see p. 490) has been interpreted as the *Zungenbecken* of the Scandinavian ice-sheet, and the Lübeck, Stettin and Danzig bays as the branch basins of the Baltic[194] (see p. 282); alternatively, the bays have been regarded as *Zungenbecken* with branch basins (*Zweigbecken*) within them (fig. 54).

Lake-zones characterise the Würm glaciation and its various retreat stages[195]—the lakes in Salzkammergut belong to the Gschnitz stage.[196] Each big Würm glacier has its terminal basin (e.g. Lake Constance of the Rhine Glacier, lakes Zürich and Lucerne of the Linth and Reuss glaciers, Brienzer See and Thuner See of the Aare Glacier, and Lac de Neuchâtel and Lake Bienne of the north-east branch of the Rhône Glacier and Lake Geneva of its south-west branch) which has "spurs" of old material converging upon it from the outer margin like spokes of a wheel.[197] The fivefold succession of the glacial series in the valleys of the Black Forest correspond to as many halts

of the ice.[198] The deep lakes of Norway (Hornindalsvatn, 514 m; Mjösa, 449 m; Tinnsjö, 460 m; Salsvatn, 445 m; Fyresvand, 369 m; Tyrifjord, 259 m; Banadakvand, 289 m) have been correlated with readvances which increased the earlier ice-erosion.[199]

It is but a logical extension to believe that rock-basins, including cirque-lakes, are lateglacial[200]; and that their distal ends fix the positions of the ice-edge at particular stages and may be used to correlate ice-margins,[201] as may rock-barriers where, as in parts of the Black Forest, moraines are lacking.[202]

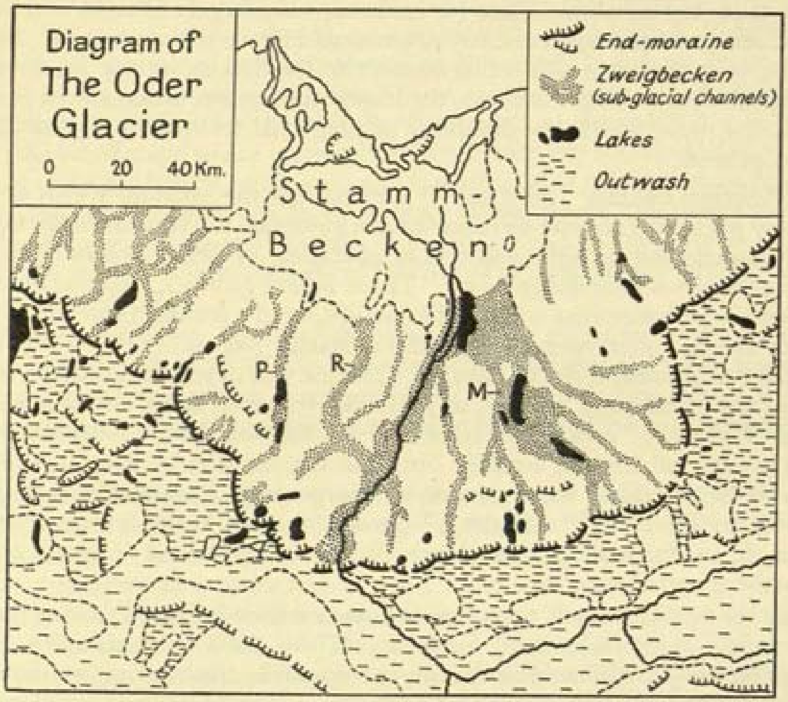

FIG. 54.—Morphological sketch of the Oder Glacier during the Pomeranian stage. P = Prenzlauer, R = Randow, M = Madü-Plöne *Zweigbecken*. P. Woldstedt, *1922*, p. 151, fig. 3.

Detraction. In a modification of the glacial theory the ice removed, either mainly or exclusively, the unevenly weathered rock,[203] rotted through immeasurable preglacial time (see p. 224) by temperature changes, moisture, chemical action and vegetation, helped by the frosts of the oncoming glaciation.[204] This "detraction theory"—the word exaration (see p. 211) has been used in this sense[205]—has gained wide support[206]: it has been specially applied to Fennoscandia,[207] the Alps,[208] North America[209] and west Greenland.[210] Many canyons and lakes in Labrador and Archaean Canada, for example, lie along the weathered zone of bands of iron ore[211] or on big, preglacially weathered dykes or suites of close-set dykes, even where these were transverse to the flow.[212]

Rotted rock unquestionably facilitated the action, especially when it was unusually deep, as along faults and joints.[213] Yet some live rock has indubitably been removed since lacustrine floors, wherever they can be

examined, are of quite sound rock. For want of crucial tests, opinions must inevitably differ as to the role played by ice and preglacial agents.

Re-excavation hypothesis. G. de Mortillet's *théorie de l'affouillement*,[214] extending the view of Venetz,[215] ascribed Alpine lake-basins to ice which ploughed out preglacial alluvium occupying pre-existing hollows provided by earlier disturbances. It reconciled the pre-existence of the hollows with the erosive powers of the ice. G. Steinmann,[216] in accepting it, substituted Pliocene marls and sands for the preglacial alluvium.

This hypothesis, also applied to south Swedish and Caucasian lakes,[217] is partly true; the subalpine lakes, for instance, were largely filled in with interglacial sediment and subsequently re-excavated by ice[218] (see p. 279). Nevertheless, it suffers as a full explanation from the defects of the preservation hypothesis (see p. 265): it leaves the basins themselves unaccounted for and awaits the proof which the discovery of preglacial freshwater deposits alone would provide.

Deflection basins. Sefström[219] noticed that the medium which striated Norway pursued a normal direction on the summits, as *Normalfurchen* evince, but was deflected to right and left of obstacles, as *Seitenfurchen* show, the latter sometimes containing lakes. These lateral grooves, likewise observed by Agassiz,[220] harmonise with Hall's conception of a frontal depression which curves like a collar round crags and obstacles (see p. 254). They are a corollary of Ramsay's glacial theory. The ice was deflected along the longer axes of the "deflection basins", as J. Geikie[221] named them, whether on land or, as Wallace's "submarine lakes",[222] on the sea-floor. These skirt the inner shores of islands, notably opposite the mouths of sounds and sea-lochs[223] with which they sometimes merge, e.g. Varangerfjord and the Norwegian Channel.[224] Geikie[225] cited numerous instances of deflection basins from Scotland as did Penck[226] from Switzerland, including lakes Neuenberg and Biel.

The ice moulded itself upon any boss over which it flowed: striae, flutings and roches moutonnées lend testimony to this. Its various layers moved simultaneously in different directions, in response to the topography, diverging on the impact side and converging in the lee. The lower ice passed over an obstacle which was low and gently sloped but split and glided past one which had a certain critical angle.[227] Thus steep-sided hills, ridges or escarpments deflected the basal layers as in the British Isles,[228] e.g. Flamborough Cliffs, Baugh Fell and Arthur's Seat, on dip and escarpment slopes near the Old Man of Wick, in the Orkney and Shetland Islands (see p. 755) and in Co. Donegal, as well as along the southern face of the Jura Moutains,[229] in the Alpine areas of transfluence and diffluence,[230] along the northern flanks of the Adirondacks,[231] and in the Finger Lakes region.[232] If no opportunity offered for horizontal or less steeply inclined motion and the obstacle was small compared with the glacier, the ice ascended the steepest slope.[233]

Drumlins and other evidence have been interpreted to mean that Edenside at maximum glaciation had a basal iceshed that bore no obvious relationship to the surface iceshed; the ice plunged downwards at these partings in an essentially fluid manner.[234] The drumlins conform to the flow of the more fluid basal ice that moved in agreement with the minor relief.

This principle of undercurrents, frequently accepted,[235] is also demonstrated by erratics,[236] by deviating osar,[237] by the absence of shelly drifts in

the Shetland Islands[238] and, except at one locality, in the Outer Hebrides,[239] by observations on existing glaciers,[240] and by experiments.[241] Striae, parallel with oblique valley lines, diverged from the normal course the ice pursued on plateaux above.[242] The narrower and deeper the valley, the more the deflection. Yet below a certain limit, the valley was essentially ignored; for a very narrow valley transverse to the flow was overridden without deflection, the ice bridging it and bearing hard on the hill tops. A broad depression, on the other hand, was crossed by ice flowing down into it and up out of it. The critical width varied with the depth and plasticity of the ice (see p. 323).

The divergence was probably associated with shear-planes (doubted or denied[243]) at the base of or within the ice, as is seen in Greenland,[244] or above spurs projecting into Alpine glaciers[245] (see pp. 117, 275). O. Ampferer[246] recognised the following shearing localities; (*a*) at inequalities in a floor; (*b*) in the lee of steps where ice overrides lower stagnant ice whose dimensions depend upon the height of the step and the velocity of the ice; (*c*) in front of and at the top of steps; (*d*) in transverse valleys where the concave thrust plane is bent according to the width of the valley and the velocity and thickness of the ice; (*e*) at lateral spurs; (*f*) at points of transfluence; and (*g*) above and below nunataks.

Although thinning, consequent upon retreat, explains many cases of divergence[247] and some of Geikie's deflection basins, e.g. that along the western side of the Minch, are tectonic,[248] an imposing residium of rock-basins athwart the main ice-flow can only be accounted for on the deflection principle.

Adverse criticisms. Many objections, most of them inconclusive or invalid, have been raised from time to time against the glacial theory as applied to rock-basins; they supplement those of a more general character urged against the theory as a whole (see p. 228). For instance, there are no basins in the beds recently abandoned by glaciers, not even at the foot of ice-cascades like that of the Rhône Glacier[249] (cf. p. 268); they are absent from the Pyrenees, Caucasus, Himalayas and Ben Macdhui and Cairngorm Mountains of Scotland[250] (cf. p. 266); they occur at embouchures where the ice splayed out and its erosion lessened (cf. p. 270); and their depth, as in a U-valley, is disproportionate to that of the ice[251] and is incompatible with a glacial origin[252] and with the modulus of cohesion of ice.[253] It is further contended that islands have persisted[254] (cf. p. 228); that ice tends rather to plane off than to emphasise inequalities by excavation[255]; and that stagnation in the form of a lens of dead ice would ensue within the hollow[256] which would stop the excavation and fill the basin with *moraine profonde*.

Flow certainly diminishes as the frictional surface increases: when this equals the shearing strength of the ice along the chord of a hollow, the basal segment will cease to move. Consequently, a definite although indeterminate limit is set to the excavation of depressions. This limit, probably as a rule not more than a moderate fraction of the ice-depth,[257] is in all likelihood rarely reached. The impediment is not the depth alone but its relation to the length of the basin.[258] This is extremely little in the only cases in dispute (see p. 263); for the ice-erosion of the small, proportionately deep lakes is generally acknowledged (see below). Moreover, ice tends to lengthen basins in the direction of flow and reduce the slopes of their floors.

That there was flow through the basins is shown experimentally[259] and by the transport of erratics across them[260]—this proof is not rigorous since the transport may have occurred earlier. Moreover, lake-floors are striated where they can be examined (see p. 266) and the negative slopes in even the deep Italian lakes were less than the gradient of the ice-surface—in Lago di Como, the least favourable case, they were respectively 30‰ and 33–40‰.[261]

Tarns. Rock-basins range from small pools amid roches moutonnées to lakes occupying hundreds of thousands of square miles. While all admit the glacial origin of the small tarns, some deny it for the bigger lakes. But, as Ramsay[262] pointed out, there is no recognisable gap between the extreme members of the series; and if the ice is capable of producing the small lake it may excavate all since depth and size are functions of duration and pressure. The argument, however, is not irreproachable; for while there is no logical halting place between the glacially eroded tarn and the largest rock-basin there is unquestionably a limit to the power of the ice in this respect. As Favre[263] has said, sand dunes do not grow to the height of the Himalayas, nor do mud volcanoes approach the altitude of Chimbarazo. It is clearly inadmissible to push to infinity an action which obviously operates on a smaller scale.

At the bottom of the scale are the "negative" water-filled hollows, underlain by rock or drift, that are strewn among the "positive" roches moutonneés (K. E. Sahlström's *ärr*[264]). Their steep or vertical face at the proximal end and gentle slope at the distal end mirror the roches moutonnées between which they lie[265]: their erosion by ice is universally recognised.

Cirque-basins. Cirque-lakes, which may be simple or may be subdivided into two or more basins, were also glacially excavated though exactly how is contested. They are an integral if not essential part of the cirque. C. Durocher early noticed that cirques and lakes coincide geographically in certain regions but erroneously regarded both as volcanic craters (see p. 295). They may be genuine rock-basins[266] or may be dammed at their lower ends by moraines which raise their level above the rock-rim. They are strikingly absent from limestones in the Pyrenees[267] and are due to solution in the Limestone Alps.[268] Some hollows in Karwendel cirque bottoms are preglacial dolines.[269]

It is agreed,[270] even by opponents of glacial erosion,[271] that cirque-lakes, the counter-parts of the *Zungenbecken* of valley glaciers, result from ice-action. Such *Hochseen* are closely connected with the glacial snowline[272] (see p. 296) and are wanting if, as in parts of the Pyrenees, Pleistocene glaciers did not exist. They are sometimes, e.g. in Tyrol, elongated in the direction of ice-flow.[273]

Cirque lake-basins were ground out at the focus of flow-lines from all parts of the cirque[274] where the pressure or stream force was greatest[275]—the *Riegel* marks the place where the stratification and flow-lines rose[276] (see p. 120). They occur, therefore, in cirques with round or armchair shapes above ridges of hard rock or where the stream precursor was constricted[277] and more rarely in valley-like extensions.[278] Material for their excavation was supplied by preglacial weathering[279] and by early glacial scree and subglacial frost[280] (see p. 301). The bergschrund too gave a steady supply of angular fragments (see p. 299) which were carried in or under the sole to grind out the rock-basins[281] (see p. 120), though they became less efficient as

by their abundance they slowed up the plastic-flowing underlayer and they became rounded towards where the basin shallows. The constant yield of similar material by plucking helps to explain why rock-basins so commonly occur at the foot of steep risers in glaciated floors. K. Gripp[282] has suggested that rock-basins, including those of cirques, owe their existence to an upward surge of advancing ice above its snout which took place because of the stau-effects of either dead ice or moraine-charged ice. Rotational slipping (see p. 118) may have played an important role.[283]

Some cirque floors may have been partly hollowed out by trunk glaciers before they dwindled back into the cirques.[284]

Eddies and rock-basins. Because of internal friction, there is no pulsatory flow in a glacier as in a stream, and eddies, generally speaking, are absent. Yet eddies, unaccompanied by turbulent flow, have been repeatedly observed in glaciers,[285] notably in the lee of lateral projections where moving ice sheared past more or less inert ice or curved back upstream. Swirls below nunataks, postulated by Suess,[286] have also been noticed.[287] The ice which sweeps between the nunataks relatively quickly is piled up to a few hundred metres on their stoss sides by the increased shear friction of the upper layers,[288] as on the Muir Glacier and Jensen's Nunataks, in the same way as water surges against the piers of a bridge. Reversed slopes, with eddies, may even exist though no rock is visible[289] but submerged ridges may be surmised.

The possibility that cirque-basins might have been hollowed out by rotating ice, in accord with their circular shape and the above observations, was raised by Richthofen.[290] Eddies were invoked for cirque-lakes in general[291] and for the Wolayer See of the Carnic Alps[292] in particular. They were called in as an aid for small rock-basins[293] including those below valley-steps.[294]

But there is no evidence of this. Not only does the great viscosity of the ice not permit it except in rare positions as mentioned above but lines of movement in glaciers are parallel and only slightly divergent. Striae on cirque-floors betray no sign of rotation.[295]

Subalpine lakes. Ascending the scale we pass through valley lakes, such as those of the Scottish Highlands and English Lake District, the glacial origin of which few dispute,[296], and reach the subalpine lakes, Rütimeyer's *Randseen*.[297] In contrast with the *Hochseen*, they lie at the foot of the Alps from north Italy to Kärnten and Steiermark. They include (with depths in brackets) Lac de Bourget (145 m), Lac d'Annecy (80·6 m), Lake Geneva (310 m), Thuner See (217·2 m), Neuenburger See (154 m), Bieler See (74·6 m), Murten See (46 m), Brienzer See (259 m), Zürich See (143 m), Lake Lucerne (214 m), Zuger See (197 m), Lake Constance (252 m), Starnberger See (123 m), Lago di Como (410 m), Lago di Garda (346 m), Lago d'Iseo (251 m) and Lago Maggiore (372 m). Some of their floors are below sea-level as in the case of the north Italian lakes—Lago Maggiore descends to −175 m, Lago di Como to −212 m, Lago di Garda to −281 m.[298] Subalpine lakes are absent from or poorly developed in the Bavarian, French and Piedmont Alps. They nestle within the Alpine valleys, as in Lake Lucerne and the Carnic Alps, e.g. in the Drau valley, or extend on to the foreland, as in Lake Geneva, Lago di Garda and Zürich See, or even lie entirely upon it as in the Ammersee and Starnberger See. These different geographical positions result in a variety of shapes; some lakes, e.g. Chiemsee, are broad and round, others, e.g.

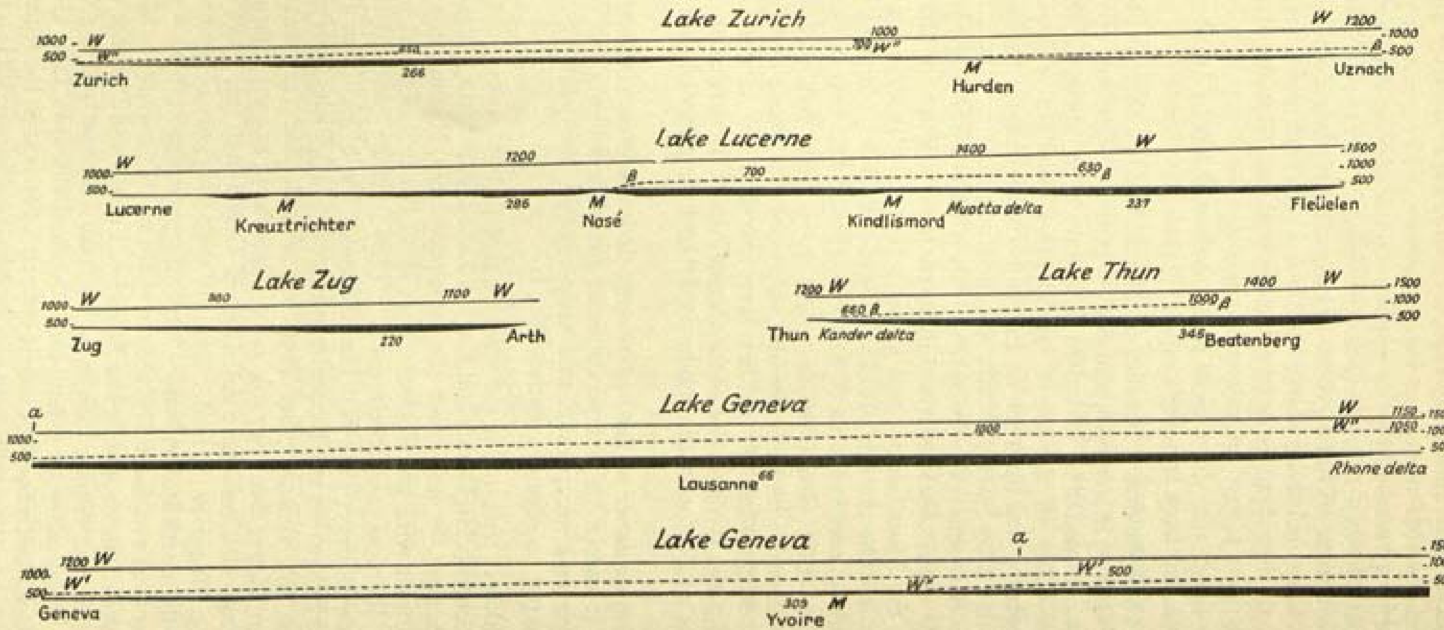

FIG. 55.—Longitudinal profiles through several subalpine lakes of Switzerland and height of the ice surface (in metres) at Würm glaciation (WW) and retreat stages (W′, W″, β). M: position of frontal moraine. The numbers beneath the lake profiles give the height of the bottom of the lake above sea-level. The profile through Lake Geneva is divided into two sections; point *a* in the upper profile must be brought to cover point *a* in the lower profile in order to achieve continuity. Scale, 1:200,000. A. Penck, *1297*, p. 596, fig. 78.

Starnberger See, are long and narrow, while others again are ramified, e.g. Lake Lucerne which is adapted to the folds in the Pre-Alps. Yet all are relatively shallow compared with their lengths (fig. 55) and betray a family resemblance and in the main have the same history. Lake Wakatipu, *c.* 378 m (1242 ft) deep, of which depth *c.* 69 m (227 ft) is below sea-level, is a good New Zealand example.[299]

It is fitting that these subalpine lakes should be the subject of the first discussion of lake origins. Nevertheless, they present the problem in its acutest form; for it is just here that the two agents which have survived critical analysis (see p. 264) attained a first-class importance. Here, if anywhere, tectonic movements, the successors of the Tertiary paroxysms, might be expected to have continued to affect relief; here too, huge Pleistocene glaciers, descending steeply from high mountains, might be supposed to have excavated severely.

With the possible exception of Lake Lucerne, the upper waters of the lakes are held up by drift dams. This is especially true of the north Italian lakes, while bores prove that the *Petit Lac* or lower basin of Lake Geneva is entirely made by a drift dam.[300] Yet the lakes are only partially dammed in this way; the true depths of some of the rock-basins have been calculated as follows[301] Lake Constance, 24 m; Zürich See, 92 m; Zuger See, 193 m; Lake Lucerne, 207 ? m; Thuner See, 155 ? m; Neuenburger See, 138 m; Lake Geneva, 255 m.

The supporters of either theory, as Heim[302] of the tectonic and Penck[303] of the glacial theory, agree that the subalpine lakes are Quaternary; as Escher[304] first observed, most of them lie within the moraines of the last glaciation. Moreover, as was also early noticed,[305] they have outwash downstream; they have interglacial deltas, as in the Garda valley[306] (see p. 1326); they coincide with glaciation (W. Kilian[307] dates Lake Geneva, Lac de Bourget and Lac d'Annecy from the Würm glaciation) and, as in the case of Lake Constance, with the flow at maximum glaciation[308] (Deecke[309] ascribed this lake to subglacial faulting of middle Pleistocene). Rhône material is absent from the Sundgau schotter.[310]

The problem is somewhat complicated by the radical changes which have affected the drainage since preglacial time. The Aare, Rhône and Rhine and other rivers between the Alps and the Jura Mountains flowed at the close of the Miocene and during the early Pliocene to the Danube[311]—the connexion is proved, for instance, by the distribution of certain molluscan species.[312] The courses of the Aare, Reuss and Limmat through the Jura Mountains were later, either antecedent to its folding[313] (the mountains were scarcely perceptible in the lower Pliocene) or initiated by backward recession across the folds.[314] Middle Pliocene gravels at the confluence of the Doubs-Sâone and Sâone-Rhine are exclusively of Rhine material.[315]

The first evidence of the Rhine north of the Jura Mountains is provided by the *gravier de Sundgau* or Upper Alsace *Deckenschotter*,[316] characterised by material from central Switzerland and the Alps. The Rhine, with the Aare and Linth, passed at that time by the Burgundy Gate and Montbéliard into the Sâone and Rhône, as suggested by M. Lugeon[317] in 1897 and since confirmed[318]; *Unio sinuatus* and other south French molluscs occur in the Mosbach sands; and the tributaries of the south Black Forest and Wasgenwald flow southwards in the mountains. Its deflection northwards from Basle,[319] due to faulting across the previous watershed at Rufach and Kaiserstuhl, followed the deposition of the Sundgau schotter and preceded the first glaciation. The absence of the Deckenschotter below Basle is due to

down-faulting[320]—a bore on the west side of the valley proved a thickness of the schotter of 230 m[321]—though Alpine erratics in the lower Rhine were extremely scarce even later when the Rhine undoubtedly flowed northwards.[322]

The upper Rhine was confluent with the Rhône[323] and the waters of the *Haut Lac* and *Grand Lac* of Geneva were captured during Günz-Mindel or Mindel-Riss interglacial times by a tributary of the Arve which drained the basin the *Petit Lac* now occupies.[324]

Glacial origin. The glacial origin of these lakes, first propounded by Ramsay,[325] has been accepted by numerous geologists,[326] including Wallace[327] who gave one of its best expositions, and has been applied to individual lakes,[328] e.g. Maggiore, Tegernsee, Chiemsee, König, Starnberg, Ammer, Constance and Plan See. It explains simply a number of geological facts and requires few complementary or accessory hypotheses. In its favour are the simple shore-lines; the few promontories or embayments; the open lake arms in submerged tributary or distributary troughs; the continuous and simple form of the slanting trough sides and their concave ascent into the triangular facets of the truncated mountain spurs; the submerged hanging junctions; and the occasional unconsumed knob and sill on the slopes.[329] The theory is reasonable if the dimensions, velocity and duration of the glaciers are borne in mind[330]—30,000 years, a fraction of glacial time (see ch. L), would suffice to give Lago Maggiore its depth of 372 m if only 2·5 cm/annum were eroded.[331] Each lake, even the different arms of a particular lake as in the case of Lago Lugano,[332] corresponds in size and depth with the dimensions of the glacier which flowed through it, and its position, relative to the mouth of the valley, with the severity of the glaciation.[333] The great depth of the north Italian as compared with the Swiss lakes is related to the larger size of the encircling moraines[334] (in part), to the steeper flexure and lower level of the erosional basis on the south side,[335] to the higher gradient and erosive power of the glaciers[336] and to the more pronounced rupture of slope at the embouchure on the plain.[337]

Nevertheless, many dismiss these arguments as inconclusive.[338] Besides the more general arguments already noticed (see pp. 228, 273), they urge that some Alpine valleys, e.g. Sesia, Brembo, Serio, Mella, Etsch, which had thick glaciers are lakeless; some lakes, e.g. Lake Geneva, lie outside the direct track of the glaciers (but *Petit Lac*, 76 m deep, impounded by drift need no longer be regarded with Penck[339] as a deflection basin); and lakes are unrelated to the character and resistance of the rocks, as glacialists assert.[340] The lakes occur where the glaciers deployed and the flow slackened and their shape, as in Lake Constance, is not what ice-erosion would produce. Finally, the north Italian lakes date back to the Pliocene[341] since beds of this age are found at the southern foot of the Alps,[342] e.g. at the south end of Lago Maggiore and in the Lugano valley, and Pliocene remanié fossils occur in drift about Lago di Como.[343] Lake Geneva is older than the earliest glaciation[344] and the lakes in general than the Kander glaciation.[345]

Tilting and drowning. Opponents maintain that while there was some erosion by ice and subglacial torrents, its amount depending upon the rate of working and duration of the Glacial period, the lakes mainly resulted from earth-movements. These lowered the Alps bodily (Ger. *Rücksinken*), thereby depressing the upper reaches of the valleys and reversing

the slope. The valley-floors were drowned and converted into elongated basins.

Although Lyell[346] suggested this view in 1864 and W. Sartorius *v.* Waltershausen[347] one year later—lakes were formed by crustal undulations over the Swiss Plain, in the Plain of Lombardy, over the Baltic and in the Laurentian region of North America—the principal advocate was Heim[348] who strongly defended it in 1878 and in subsequent years. He cited the backward tilt of the preglacial surfaces, Deckenschotter, moraines and rock-terraces, as well as the submerged terraces and increasing thickness of the accumulations which floor the valleys as the Alps are approached or entered so that the rocks are concealed except in epigenetic courses. As confirmation have been mentioned the backward tilt of the Molasse near Berne,[349] of the Deckenschotter about Zürich See,[350] and of the moraines of Lago d'Iseo[351]; the great depths of the lakes near their heads; the lacustrine outlines which are just those that subsidence would impose[352] (see below); the flooded lateral valleys, as around Lago di Garda, which Penck[353] admitted; the absence or occurrence only at considerable depths of lake-dwellings in the Alpine parts of Zürich See, Lake Constance and Lake Geneva because of post-glacial tilting and subsidence[354]; the postglacial movements of Ammersee and Chiemsee; and the "mass deficiency" of the Alps.[355]

The backward slope of the terrace lines and remains of old valley floors, first noticed on Zürich See by Heim, has since been affirmed for this lake by others[356] and stated for other lakes,[357] e.g. Lake Constance, Lake Geneva, Thuner See, Brienzer See, Lac de Bourget, Starnberger See, the lakes of Neuchâtel, and for Lago d'Iseo and Lago di Como on the south. The gradients of the terraces, stated to be 15–25‰,[358] give with the original outward slope of 5–8‰, a tilting of 22–30‰ or a vertical subsidence of 120 m.

Heim[359] was of the opinion that the Alpine valleys had already reached their present depths during the great Mindel-Riss interglacial and before the High Terrace schotter filled the valleys. Some erosionists[360] have also held that the valleys were eroded to their present depths during the oldest glaciation. The subsidence, which amounted to 200–300 m on the north and 300–500 m on the south,[361] was fully accomplished at the end of the great interglacial epoch since it did not affect the moraines of the last glaciations. Indeed, but for these, the valleys would now be filled with sediment[362]—a partial return to the preservation theory (see p. 265). Heim gave the extent of the interglacial lakes[363]; on the south, fjords opened into the sea of the Po basin (see above).

Several of these interglacial lakes have now been established[364]; there are interglacial deltas, for instance, in Lakes Zug, Zürich and Thun and in the valley of the Tagliamento. Their occurrence at levels above the modern lakes, ascribed by Penck to morainic dams, is best explained by interglacial subsidences, as recognised for Alaska[365] and for the Alps by O. Ampferer and later by Penck himself (see p. 446).

Heim ascribed the depression to a marginal flexure, parallel with the Alps from Bavaria to Savoy. Others have resurrected Heim's earlier view[366]; they seek the tilting not in a backward sinking of the Alps, but wholly or partly in a rise of the extra-alpine foreland, including the Jura Mountains and the Molasse anticlines.[367]

While Heim and others imagine the Alps sank *en bloc* as a mechanical unit, unaccompanied by differential folding and faulting, Ampferer from studies

in the Inn valley concluded that the subsidence varied locally (see p. 446). Thus the schotter near Rum, 200 m thick (a further 300 m occurs above the valley-floor) shows no upward shallowing but rather a number of superimposed *Verlandungsserien*. Penck and Ampferer imagine a synclinal bending with a deepest point above Innsbruck.

The subsidences may have been tectonic and not glacio-isostatic,[368] although their maxima were under the thickest ice.[369] The cause, however, is still uncertain and disputed (see p. 446). Heim[370] sought it in an isostatic depression of the Alps, overloading following a piling up of the beds during the Tertiary era. He envisaged it as the final stage of mountain building or a tectonic sinking in a geosyncline[371] that was still continuing.[372] The fact that the Caucasus had no subsidence he linked with young eruptive activity. Others[373] have thought the Alps were glacio-isostatically depressed and have not yet fully recoiled, though this is difficult to reconcile with the lacustrine depths and with the probable isostatic lag.[374] Gravity determinations show an excess in the Alps which may be due to incomplete compensation, either because the movements are recent[375] or because of the support of the surrounding land.[376] The Swiss lakes, on the other land, lie in a zone of gravity deficiency.[377]

Although the theory of tilting and axial subsidence for the Alps has received substantial support[378] and has been applied elsewhere, including the lakes and fjord-basins of the North American Cordillera,[379] it has been severely criticised by glacial erosionists ever since Ramsay[380] himself first opposed it. Apart from more general arguments, such as the composite nature of the lakes and their coincidence with glaciation, Wallace[381] rejected it since their size, depth and position derived no meaning from it and their outlines and straight sides and their subaqueous contours, devoid of submerged channels, militated against it. With the extremely slow tilting, the barriers would be sawn through before the levels were raised and the oversteepened slopes would be stripped of their waste while the lessened slopes would retain theirs.[382] Furthermore, valleys without lakes occur between valleys with lakes, e.g. the Mella valley between Lago d'Iseo and Lago di Garda[383]; and the valley-floors of the earlier cycles (see p. 324), the Deckenschotter and later terraces rise towards the Alps[384] (see p. 1326). Equally inimical is the position of the lakes relative to the valleys and foreland,[385] to the geological structures[386] and to the axis as defined by Heim which, north of the Swiss Alps, runs through the north part of Zuger See and the longitudinal Reuss valley below Luzern and through the upper half of Lake Geneva.[387] The latter lake is independent of tectonic structure.[388] Finally, as Heim[389] recognised, the theory is unable to explain features like rock-barriers, *Inselberge* and steps upon which the glacial erosionist lays much stress.

The gravest objection is taken to the affirmation that the terraces slope backward. Those margining Zürich See are solid features differentially eroded by ice; they coincide in dip and strike with the beds themselves and multiply up the valley as the hard beds themselves do.[390] The single tilted Deckenschotter is really three horizons[391]; the terraces coincide in places with conglomerates in the Molasse[392] and cannot be due to backward tilting in Lake Geneva and Lake Constance which are situated outside the Alps. The features are absent from Zuger See, Lake Geneva and Lago Maggiore.[393]

The schotter and interglacial deltas (see above) prove upheaval rather than

subsidence, as do the doming of the *Gipfelflur*[394] (see p. 327) and the geological history of the Alps since the earliest Miocene.

An uplifted foreland is similarly irreconcilable with the low altitude of the floors of Lake Geneva and the north Italian lakes,[395] and with the slight change in Quaternary levels north of the Alps as shown by terraces in the Danube gorge[396]—the post-Tertiary cutting at Passau is estimated at 25–30 m only.[397]

From this discussion it may be gathered that it is still too early to make a positive statement of the history of the *Randseen*. A wise working hypothesis should be sufficiently comprehensive to embrace tectonic movements, which have operated for reasons only dimly perceived, and also changes linked with multiple glaciation and interglacial fluviatile phases that cannot yet be fully reconstructed.

Baltic Sea. The Baltic Sea, whose geological history W. Deecke[398] has sketched, lies principally in the Scandinavian–Russian platform but extends into the Archaean and Hercynian regions. It has a maximum depth of 261·89 m and forms part of the great geosyncline, between the Scandinavian Shield and the mountains of central Germany, which has been a lasting feature in the face of Europe, even from the Lower Palaeozoic,[399] and was completely filled in with Palaeozoic sediments. The sub-Cambrian surface of denudation, which is fairly high in Norway, covering much of the Hardangervidda and the central unbroken block,[400] lies at −387 m in Gotland, at −200 m in Estonia, and at −180 to −190 m at Leningrad. Although the history of the warping (De Geer's Baltic Flexure) is imperfectly known, its continuation into the Tertiary may be postulated from the sediments of this era in Denmark and north Germany.[401]

A great basin or lowland, developed in connexion with the late-Tertiary Scandinavian uplift, occupied this site in the upper Tertiary[402] and at the beginning of Quaternary time.[403] Wide open valleys in the peneplain about the Gulf of Bothnia drained centripetally into it,[404] and large river-valleys, now filled with Tertiary detritus,[405] traversed the Cretaceous rocks from Ystad to Åkarp between Malmö and Sund. Holst's *Alnarpsflod*,[406] with its wood, fruit, leaves, mosses, molluscs, insects and amber (*Bärnstensflod*), flowed south-eastwards (north-westwards?) along a valley (possibly tectonic[407]) that runs for at least 7 km and descends to −60 m and provides the largest source of underground water in Sweden[408]; the channels at the bottom of the present outlets of the Baltic were probably excavated later by subglacial streams or by overflow waters from glacier lakes (see p. 1290). The *Alnarpsflod* was apparently merely one of many Pliocene streams which carried their sands and gravels southwards into East Prussia and Poland.[409]

It is almost certain that while a basin existed on this site during the early Pleistocene—it guided the Early Baltic Glacier (see p. 711)—there was no sea.[410] The northern part was then dry land and much higher, as it was throughout most of geological time.[411] Even north Germany and the Belts were above the sea. The strandflat sea did not enter the basin, though its occurrence is affirmed[412]—its non-detection in the Gulf of Bothnia has been doubtfully referred to later submergence[413]—and the ice found no marine shells to carry into the drifts of north-west Russia, as G. v. Helmersen observed, or of Mecklenburg,[414] nor Pliocene shells into the drifts of north Germany and north Europe.[415] During the Pliocene, the Baltoscandian rocks were undergoing erosion.[416]

The earliest marine incursions were the Holstein and Eemian Seas[417] (see

p. 944) and only those,[418] including the monoglacialists, who relegate these clays to a preglacial instead of the interglacial date generally accepted, believe that the Baltic Sea antedated the Ice Age.

Although opinion is virtually unanimous about the age of the Baltic Sea, it is probably more sharply divided on the question of glacial erosion *versus* tectonic movement than in the case of the subalpine lakes. The preglacial lowland was certainly modified by glacial erosion, by deposition and by movement. Yet views differ profoundly about the modification each made and particularly that which the ice produced. Many hold that the basin was largely or entirely ice-eroded[419]; it is the central basin of the Scandinavian ice-sheet[420] which pressed the Baltic Ridge up before it as a solid wave,[421] excavated the radially arranged peripheral depressions, the föhrdes of Jutland and Schleswig-Holstein[422] (see p. 353) and the north German *Rinnenseen* (see p. 241) and bays,[423] e.g. in Mecklenburg and Pomerania and of the Oder and Vistula (see p. 270), and deposited the material as the Baltic moraines,[424] etc.

Glacial erosion on a vast scale is demonstrated in many ways. Numberless erratics in the circumbaltic belt are referable to sub-Baltic sources (see p. 369)—the Baltic's floor provided the bulk of the igneous and all the sedimentary erratics[425] such as flint, chalk, Silurian limestone, Kellaway, Kimmeridge, Portland, Neocomian and Albian in the Pleistocene deposits of Denmark.[426] The north German drift has a high percentage (80%) of quartz-sand (from the Miocene strata at the bottom of the Baltic), compared with the Swedish drift[427] (31%), its boulder-clay is rich in lime[428] derived from the chalk, and its sands and gravels have abundant flint.[429] Chalk preponderates in the upper drifts of East Prussia[430] because the Tertiary cover was stripped off during the earlier glaciation.[431] The *Zungenbecken* at the mouth of the Vistula was formed by the progressive erosion as proved by the *Schollen* in the drifts to the south (see p. 363), viz. the Miocene at greatest distance, Oligocene somewhat nearer and finally Cretaceous near the delta.[432] The whole of the Tertiary strata and the upper part of the Chalk were removed from north-east Denmark,[433] as were the Cambro-Silurian formations from the Swedish Baltic islands[434]; and the northern boundaries of these and other formations were displaced southwards. The Baltic widens and shallows on the Silurian, Devonian, Mesozoic and other soft rocks and narrows and deepens over hard rocks like granite and porphyry.[435]

Yet tectonic movements have been active. Powerful dislocations, partly of Mesozoic age, with downthrows in places of several thousands of metres, are known to run along the west coast of Sweden into Scania[436] and Bornholm[437] (Tornquist Line[438]) and along the Pomeranian coast,[439] while the fault troughs of Scania, the Stockholm archipelago and Rügen continue as traceable depressions on the floor of the Baltic.[440] This, indeed, like the central Swedish lake-region, is broken up into blocks, some of which, e.g. Bornholm, remain as horsts while others have sunk, as in the case of the Bornholm Deep: the chalk stands at different levels. A trough, *c.* 300 m deep, skirts the west coast of the Gulf of Bothnia for *c.* 60 km off the towns of Härnosänd and Örnsköldsvik, or almost exactly opposite the area which was most uplifted epiglacially[441] (see p. 1315). Other Baltic "deeps" have been linked with postglacial movements.[442]

Tertiary faulting, contemporaneous with that which initiated the Norwegian

Channel and Scania's dislocations, affected the Baltic in general,[443] and the gulfs of Bothnia[444] and Finland,[445] the south Baltic[446] and the plains about Möen and Rügen[447] in particular. It determined the rhombic outline of the Bornholm horst[448] and affected lakes Onega and Ladoga and the White Sea,[449] Skagerrak and Kattegat[450]—the "Danish Strait" of the Belts and Jutland is a depression probably as old as the Pompeckj rise (*Schwelle*) of the Elbe region.[451]

The age of the faulting is less certain. It was definitely Tertiary in the main but may have revived posthumously during the Pleistocene[452] as has been asserted for Möen and Rügen (see p. 257) and demanded as a corollary of the great erratic *Schollen* (see p. 364). It may have continued into post-glacial time.[453]

The foregoing suggests a complex origin for the Baltic in accord with its complex shape. Renewed flexuring, faulting of considerable magnitude and varying regional intensity, together with glacial erosion on no mean scale, have played their parts. To assign to each its relative role is a task for the future.

Finger Lakes. The Finger Lakes of New York, twelve in number and up to 40 miles (*c.* 65 km) long, notch the nearly horizontal strata of the northern edge of the Appalachian Plateau on a plan suggesting fingers spread apart and pointing southwards (fig. 56). Their origin has provoked much controversy. The antithesis, however, is not glacial erosion and tectonics, as in the European cases just examined, but glacial erosion and glacial deposition. One school[454] holds that the ice-flood heightened the preglacial relief and eroded the valleys through a vertical amount estimated at 1500 ft (*c.* 460 m) and gave them their depth, their smooth U-shaped contours and their straight boundaries; for they are parallel with the ice-flow, are deepest in their narrowest parts, and the lowest parts of their rock-bound tributaries, now filled with drift, are above the present lake-surfaces—both the Seneca and Cayuga lakes have tributaries that hang above the main valley-floors by 400 ft (*c.* 120 m). The valley-floors are excavated in shales that dip gently southwards and presented optimum conditions for plucking.

According to the second school,[455] the valleys were rejuvenated from the north in Tertiary time and were converted into lakes by the extremely thick drift of the submarginal zone, including the great drumlin belt (see p. 393). Fairchild,[456] its protagonist, states that preglacially weathered rock and interglacial gorges are still intact; the drift, as revealed by bores, is very deep—in parts of the Seneca and Onondaga valleys it is well over 1000 ft (300 m); and the ice stagnated in the valleys and bay-like indentations south of Lake Ontario that overlie the buried valleys.

So far as is known, the lakes are ponded entirely by morainic deposits, since rock-basins have not been proved to exist. East and west from the two largest lakes (Cayuga and Seneca), the lakes increase in altitude to almost 900 ft (275 m) while decreasing in depth and size. Lake Cayuga, which is 381 ft (116 m) above the sea, has its bottom 54 ft (16·5 m) B.S.L. and Lake Seneca, 440 ft (134 m) above the sea, reaches a depth of 174 ft (53 m) B.S.L. A bore 600 ft (183 m) below the bottom of this lake failed to reach rock-floor. The floor of the Onondaga valley is also below sea-level.

Great Lakes. The Great Lakes have occasioned much debate and a

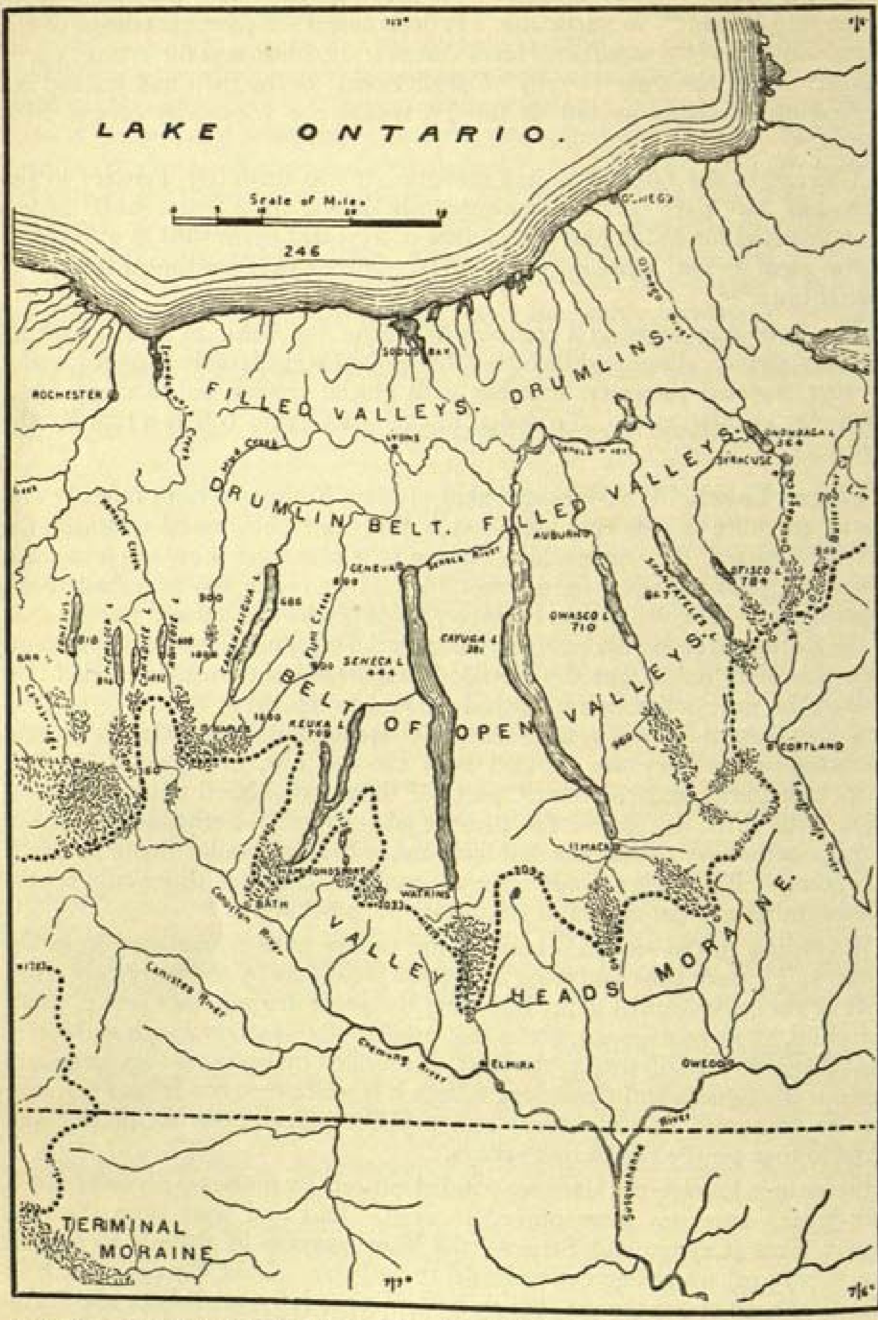

FIG. 56.—Map of the Finger Lakes region, New York. H. L. Fairchild, *P. Roch. Ac.* 6, 1929, pl. 88.

voluminous literature (cf. bibliographies[457]). Their dimensions, as given by the U.S. Coast and Geodetic Survey, are as follows:

Lake	*Surface area (sq. miles)*	*Height of surface A.S.L. (ft)*	*Greatest depth (ft)*	*Relation to sea-level (ft)*
Superior	31,820	602	1290	−688
Michigan	22,400	581	923	−342
Huron	23,010	581	750	−169
Ontario	7,540	246	774	−528
Erie	9,960	571	204	+367
Nipigon	1,769	850	402	+448

The lakes are essentially basins in undisturbed strata though Lake Superior, like the shallow Lake Nipigon, lies in an ancient synclinal structure in the Keewenawan rocks and a fault trough[458]; Lake Michigan and Lake Huron repose in a monocline of the soft rocks overlying the Niagara Limestone—Lake Erie rests in a trough on the same horizon and Lake Ontario in a monocline beneath that limestone. The escarpment overlooking Lake Erie and Lake Ontario demonstrates their origin by excavation.

All these lakes lie in soft rocks, either Ordovician (Green Bay, Georgian Bay, Lake Ontario) or Devonian (Lake Michigan, Lake Huron, Lake Erie).

The glacial erosion theory was extended to these lakes by Ramsay,[459] Hind (see p. 211) and many later geologists.[460] Penck,[461] for example, likened them to *Zungenbecken* in their relation to the morainic belts and Martin,[462] in computing the overdeepening of Lake Michigan at 500–800 ft (152–244 m) and of Lake Superior at 600–900 ft (183–274 m), gave the view quantitative expression. In its application to the Great Lakes, the glacial theory in its more reasonable proportions has attained its severest form. Like lakes Ladoga and Onega of the Glint of Europe and Great Slave Lake and Bear Lake (respectively 826 ft and 450 ft or 252 m and 137 m deep) and other glint lakes to Coronation Gulf on the western edge of the Canadian Shield, they were eroded by ice-sheets; the earlier drifts have been occasionally reached by erosion[463] or penetrated in wells.[464]

In justification, it is said that there was no preglacial lake on this site (a belief all workers share) and no preglacial valleys around Lake Superior; the ice-flow was independent of the escarpments; ice-scratches pass into and out of the lakes; the sides are relatively steep and the floors irregular and at great depth—the deepest part of Lake Ontario lies directly north of the Finger Lakes depression where the escarpment is all but absent; the drifts to the south are of great thickness (see below); similar basins occur in other glaciated regions; and the difficulties in accounting for the basins in any other way are considerable.

Against this view, which perhaps inevitably rests upon very little positive evidence (see p. 265)—the third argument is not valid (see p. 266) and the first does not rule out other agencies—it is objected that the summit edge of the Niagara Limestone is still fairly sharp, glaciation not having planed it off[465]; the Wisconsin ice was lobar, proving that the basins then existed (but may have been eroded by earlier ice-sheets); there are islands of weak rock in the Lake Superior basin; and the drifts on the south, if replaced, would fill only a small fraction of the hollows.[466]

Some hold that the relief resulted preglacially[467] when a river-system drained the region. Many writers, e.g. A. P. Coleman, T. C. Chamberlin, E. W. Claypole, F. Leverett, R. D. Salisbury, J. W. Spencer and I. C. White,

have thought that the country now watered by the upper Mississippi, Ohio and Susquehanna then drained northwards. Thus according to Spencer,[468] an immense Laurentian River, a predecessor of the St. Lawrence River, flowed from the site of Lake Michigan across Lake Huron, down Georgian Bay, along a channel, now filled with drift, into the Lake Ontario area (fig. 57). In favour of this plan are the adjustment of these master streams to

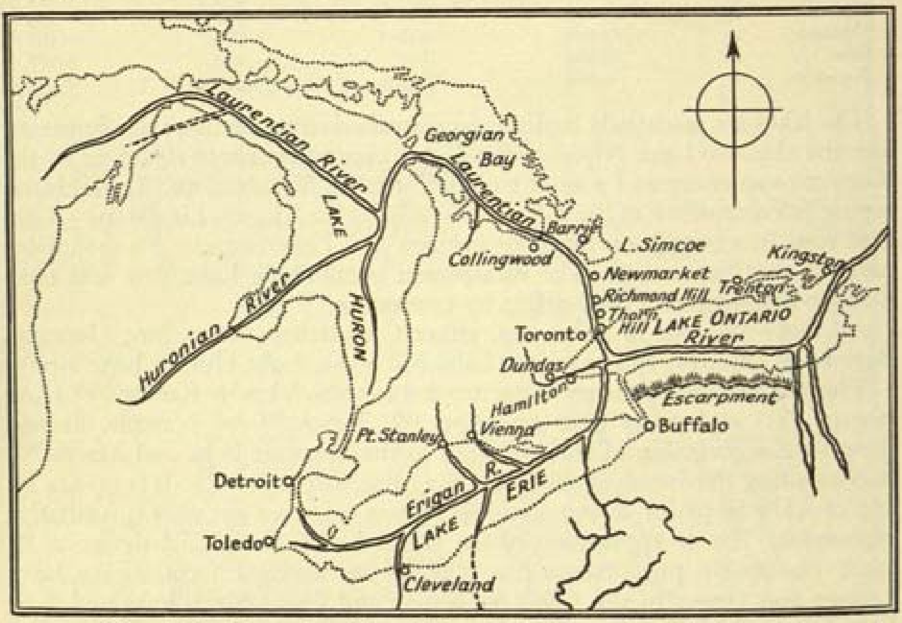

FIG. 57.—Map of the Laurentian River (After J. W. Spencer). The Erigan River probably entered the Erigan valley west of Hamilton (A. P. Coleman). *298*, p. 11, fig. 3.

structure and their confluences at not improbable angles. The Mississippi, then without the Missouri (which the Illinoian drift diverted from the Nelson drainage) was much smaller than now. Others[469] imagine a reverse drainage into the Mississippi which was turned northwards by earth-movements. A. W. Grabau's reconstruction[470] depicts three ancient rivers, the Saginaw, Dundas and Genessee rivers, all guided by joint or fault lines[471] and trending south-west as consequent streams on the slopes of the Tertiary peneplain. Yet a third view[472] interprets these valleys as glacial channels margining a lobe of ice issuing from the area to the north-east (see p. 974) and which carried many times as much water as the present River Mississippi.

This preglacial relief (it may be interglacial), which A. P. Coleman[473] has established in some parts, was modified,[474] occasionally by faulting, more commonly, as about Lake Erie and Lake Ontario, by warping attributable to or independent of glacial isostasy. It has also been appreciably altered, notably in the case of Lake Michigan and Lake Huron (including Georgian Bay), by filling in the old river-beds with drift and diverting the drainage along epigenetic courses over solid rock so that large tracts were drowned.[475] Spencer[476] in 1881 found the buried preglacial outlet (Dundas valley) of the Lake Erie drainage into the west end of Lake Ontario, its floor descending to more than 470 ft (143 m) below the surface of the lake. Other cases of buried valleys connecting lake-regions or differing from the present river-courses

have been described,[477] e.g. that between Lake Huron (Georgian Bay) and Lake Ontario which descends from the surface at 650 ft (198 m) to sea-level or 246 ft (75 m) below Lake Ontario; that in the north part of New York State; that between Lake Superior and Lake Huron, west of the present connexion by St. Mary's River; and that between Lake Michigan and Lake Huron through the Mackinac Strait. The inundation has submerged escarpments around Lake Michigan and channels in lakes Huron, Michigan and Erie as pointed out by S. Hunt and J. S. Newberry. Its effects are also seen in the drift-filled valleys, locally 400–500 ft (122–152 m) deep, which fall into the lake-basins.[478] Similar steep-walled valleys, filled with drift to a depth of 495 ft (151 m) and tributary probable to the Nelson, are also known from north Manitoba. The upper Missouri and Yellowstone rivers may have flowed originally towards Hudson Bay, joining the greatly lengthened Nelson.[479] The continental divide between the Arctic and the Gulf of Mexico has been shifted far to the south.

Hence the Great Lakes must be ascribed to the joint action of ice-erosion, the exact magnitude of which is debatable, of glacial deposition of whose existence there are opportunities of positive knowledge, and of crustal movements. The hydrographic depression of the preglacial rivers which may have drained northwards has now been modified and in some degree obscured by ponding following the closing of the ancient valleys by crustal movement and by the deposition of thick drifts. Ice-erosion of uncertain amount has added to the complexity.

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Chapter XIII

CIRQUES

Form. Cirques—J. de Charpentier[1] was apparently the first to recognise the type and introduce the word—are hemispherical bowls. The encircling cliffs, usually steeper behind than at the sides and sometimes unscalable, often sweep up to narrow knife-edges or sharply serrated crests suspended like garlands between the peaks. They frequently rise 300–500 m from the floor; in the Hohe Tatra they average 533 m[2] and in the east Tauern 670 m[3]; the rear wall of the Walcott cirque in the Antarctic is about 10,000 ft (3000 m) high.[4] The slope varies with the lithology; it is sometimes sheer in hard, resistant rocks and is steeper, for example, in the Limestone than in the Central Alps.[5]

These cliff-bound stadia ("armchair" shaped hollows[6]) are semicircular or horse-shoe shaped in plan, though spurs may partition them into twin or multiple theatres. The transverse profile is generally U-shaped, especially in limestones,[7] but is often funnel-shaped[8] (Ger. *Kartrichter*). In the ideal form, a lake bounded on its outer side by an arcuate moraine and replaced sometimes by peat, nestles in the bottom (see p. 274); these lakes are the tarns of the English Lake District, the *Meeraugen* of the Tatra[9] and the *cocha* of the Andes of Peru and Ecuador.[10] The cirque-lake may completely submerge the cirque-floor or cover only a fraction of it. While the floor in these cases inclines towards the back of the cirque[11] (Ger. *Rücktiefung*), particularly in steep-walled cirques[12] as in the eastern Alps, in some, notably in the highest and steepest mountains, e.g. the central Alps,[13] it falls sharply outwards to the main valley—the two types have been termed "closed" and "open" cirques.[14]

Cirques occur on glacial divides as (1) hanging cirques (Ger. *Gehängekare*[15]; Norw. *Fjeldbotn*[16]) which, perched on mountain sides lateral to the major valleys, become fewer as we penetrate the mountains[17] or (2) terminal or valley-head cirques, at the upper ends of main valleys (Norw. *saekkedal*; *blinddal*; Helland's *dalbotn*; A. Aigner's *Talschlusskar*; Richter's *Saektal*)[18]. Some would restrict the term cirque to hollows in the first position,[19] others to those in the second,[20] a distinction difficult to apply and less preferable to general usage which agrees with etymology as well as with geographical convenience. The suggestion that it should embrace all similar forms of whatever genesis[21] has little to commend it.

Cirques are cradled in almost all kinds of rock. In Scotland, for instance, they scar Pre-Cambrian schists, etc. in many places including the Shetland Islands and Harris, granite in Skye, Arran, Galloway and the Cairngorm Mountains, gabbro, dolerite and peridotite in Skye, Torridonian Sandstone in the north-west, Silurian slates and grits in Galloway and Old Red Sandstone in the Orkney Islands. In Snowdon, they are in various kinds of igneous rock.[22] In Ireland, they are hewn in granite in the Donegal, Mourne and Wicklow Mountains, in slates and grits in the south and in Dalradian rocks in the west. While they are mainly associated with granite in the Tatra,[23] they are combined with all varieties of rock in the Alps[24] and in

Norway[25] (though mostly with the Archaean). In the Pyrenees,[26] they are situated in granite and crystalline slate and in the Apennines[27] in Mesozoic and Tertiary limestones and sandstones (pl. X A and B, facing p. 257).

Although this suggests that cirques are independent of structure and petrological character,[28] it is not so since the jointing, cleavage and direction of dip control both size and shape.[29] Jointing is generally the more important unless the stratification be highly inclined; for walls and joint systems are often parallel. Regular and simple shapes occur in strata of homogeneous texture and equal hardness or in those which are horizontal or dip into a hill. Where joint, bedding or foliation planes slope outwards, the floors tend to fall outwards too and the head walls to coincide with joints and generate slides: the lateral walls are then very steep. Cirques in the Black Forest are confined to a definite layer of the *Buntsandstein*[30] and as in the Vosges to horizons where hard and soft strata meet.[31] In the Carnic Alps, they also frequently mark the contact of different kinds of rock[32] and in the Riesengebirge the junction of harder and softer (porphyritic) granites.[33]

Four conditions tend toward perfection of cirques[34]: preglacial valleys, sufficiently widely spaced to allow development without mutual interference; climatic conditions not so severe as to cause glaciation of the intervening divides and uplands; fairly homogeneous rocks; and little postglacial crumbling and erosion.

Pseudo-cirques. Although cirques are widely distributed in all glaciated lands (see below), similar but not homologous forms are to be seen in unglaciated countries. These "pseudo-cirques"[35] have been described, for instance, from arid areas, such as the Arabian and African deserts,[36] including the limestone desert of Lower Egypt,[37] and the plateau above the Colorado Canyon,[38] i.e. territories in which there is no downwash and where wind removes the debris resulting from insolation. They are simulated too in limestones,[39] as in Swabia, Salzkammergut and the Limestone Alps, and in horizontal strata or rocks with pronounced jointing like the *Quadersandstein* of Sächsische Schweiz,[40] the *Buntsandstein* of south Germany,[41] soft rocks in parts of Asia Minor,[42] or sheet-jointed granites in Sinai.[43] They are especially apt to arise from landslipping,[44] as in the north Limestone Alps, Big Snowy Mountains, Montana, and the Pennine Chain. Some cirques may be unaltered preglacial landslip hollows[45] while ice may have modified others.[46] Landslipping, by aiding a glacier's tug, may indeed under suitable conditions help cirque-formation.[47]

Non-glacial origin. Cirques were early misinterpreted as volcanic craters[48] and even later had a volcanic origin assigned to them,[49] as in central France, Spain, the Andes of Ecuador, and Ross Archipelago, Antarctica. Solution has also been invoked for those in limestones[50] and others, as in the Balkans,[51] have been regarded as glacially modified dolines.

In the days when the belief in glacial submergence held sway, cirques were attributed to marine erosion.[52] Deformation or tectonic subsidence was also called in,[53] notably in the Vosges, Pyrenees and Bohemian Forest. But cirques are restricted to definite zones of altitude (see below), are absent from many faulted terrains, and occur where no dislocation is detectable.

T. G. Bonney[54] examined cirques in calcareous and other strata of the Alps and the crystalline rocks of Skye. In his view, shared more or less by a number of other geologists,[55] they were excavated preglacially by small convergent streams working backwards into the hills. But water erodes linearly

and does not raise thresholds. Nor is it present in any quantity in the firn region in which cirques are typically developed.[56]

Glacial erosion. In view of the weakness of its rivals, it is not surprising to find the glacial origin of cirques, first propounded in 1860 by Ramsay,[57] applied during the following quarter of a century in most parts of the world[58] and almost universally in recent time.[59] Positive supporting evidence is overwhelming though workers during the 19th century did little more than appeal to the theory's reasonableness. Cirques, as Ramsay observed, are intimately related in their distribution to present or former glacierisations: they are the "characteristic fossils" of ancient glaciers[60] and are found throughout all glaciated mountains (though weakly developed in the Caucasus[61]) as their numerous local names[62] reveal: *Kar*, Bavaria and Austria (of interesting derivation[63]); *Grube*, Riesengebirge and Tyrol[64]; *cwm*, Wales; *combe*, English Lake District; *combe* (*cum*) and *lug*, Ireland; *caldare*, Rumanian Carpathians; *zanoga*, Slav mountains; *rupe*, Balkans; *oule*, Pyrenees; *coire* (*corrie*), Scotland; *botn*, Norway; *hvilft*, Iceland; *circo*, Italy; and *corral* and *hoyo*, Spain. These names are equivalents, though it has been suggested that cwm or corrie should be restricted to the individual high mountain valley, the term cirque being applied to the larger, often composite feature of the great alpine regions of the world.[65]

Cirques and snowline. Cirques, as F. Ratzel noted and Penck's table[66] brought out, encircle mountain flanks at a definite height peculiar to each group. The girdle runs parallel with other altitude zones and is closely related to the glacial snowline, as has been demonstrated for many regions,[67] e.g. Norway, Greece, the Alps, Pyrenees, Apennines and Andes. It has been frequently used, as by Geikie[68] in Scotland and by Penck and Richter in the Alps[69] and more recently in the western United States,[70] to fix the height of this line. The lighter the snowfall, the closer is the relationship.[71]

Thus the cirques rise from higher to lower latitudes,[72] e.g. north-west Iceland, 200–400 m; central Norway, 1000–1600 m; Himalayas, 4000–5000 m; Peruvian Andes, 4300–4600 m; Patagonia, 1000 m; New Zealand, 600–1200 m. In the Lofoten Islands, north Fennoscandia, north Labrador, parts of Greenland and Alaska, Tierra del Fuego, and Antarctica, e.g. South Georgia, they are at or below sea-level[73] and give rise to fjords of the specially broad, half-oval type; Spitsbergen's alpine massifs are sunk into the sea up to the ends of the glaciers. Cirques also ascend from the margins into the interior of mountains,[74] as in the Hohe Tatra, Pyrenees, Vosges and Scottish Highlands, and rise eastwards in Europe as the following data bear out[75]: Vosges, 570 m; Black Forest, 750 m; Bohemian Forest, 930 m; Riesengebirge, 1030 m; Altvatergebirge, 1200 m; Hohe Tatra, 1500 m.

A corollary is that the tiers of cirques (see below) in some mountain groups coincide with the successive positions of the snowline in its upward retreat.[76] Thus de Martonne's "parasitic cirques"[77] of the higher Carpathians belong to the rising snowline of the later and smaller glaciers, as do the *Schneegrenzkare* of Fels[78] which also flank the mountains in tiers.

The optimum conditions for cirque development occur apparently (and for reasons to be analysed below) in a zone in which the average annual temperature is somewhere near freezing point.[79] Where the temperature is too low, for example at sea-level on parts of the Antarctic coast to-day, the process is

virtually at a standstill—this cirque excavation took place in a milder period before the present frigid stagnation set in.[80]

A genetic connexion with snowlines has sometimes been questioned.[81] Many Alpine cirques were buried under the Pleistocene ice—their floors were 100–200 m below the ice surface and several 100 m above the snowline of maximum glaciation[82]; and the altitude of the floors is highly irregular[83] (this is partly owing to aspect and to the glacier's duration and erosive powers) and often related to preglacial relief,[84] e.g. rejuvenated heads. Other workers[85] admit the relationship but affirm that cirques are unsuited to determine the snowline. Although observations[86] in the Tauern and Vosges prove that caution is necessary in using this method, it is nevertheless true that small cirques, particularly if at the glaciated periphery, are well adapted for the purpose[87] and for checking snowlines obtained by comparing the lowest cirque levels with the altitude of unmodified gullies.

Tandem cirques. Cirques in almost all glaciated regions frequently occur in tiers,[88] the "tandem cirque" or "cirque stairway"[89] (Ger. *Kartreppe*: Fr. *escalier de cirque*), each step often with its rock-basin and tarn. The heights of neighbouring steps in any one region are practically the same, as stressed by Penck,[90] who stated, somewhat hesitantly, that each step is related to convergence or to a position of the ice during its lateglacial retreat. This second relationship, applied by others to various glaciated centres[91] (Alps, Pyrenees, Altai, Andes, New Zealand) and to steps in hanging valleys,[92] was extended by other glacialists[93] who equated each step with a glaciation as Hess[94] did on other grounds. Paired lakes, separated by a vertical interval of 90 m, have suggested that the Yolande glaciation of Tasmania (see p. 980) had two phases.[95] Complications are provided by reconstructed glaciers which are formed on lower shelves or niches out of avalanched ice descending from higher levels, as well as by structural flaws in the rocks which produce ledges on which small névés or glaciers, by nivation and ice-erosion, may dig themselves in to make corries of an upper storey.[96] Interglacial stream action is also thought to have played a part (J. Blache, 1952).

The connexion is denied[97] because the stages were too short-lived, the steps in different cirques are disharmonious in altitude, and stadial positions do not coincide with cirque-thresholds. Others[98] again who share these objections see in the tiers evidence of successive, interglacial uplifts: Richter[99] had previously interpreted the steps as preglacial features accentuated by ice. This conclusion finds acceptance and support[100] in the transition from tandem cirque into small stepped hanging valley, analogous with the steps in U-valleys (see p. 328) and already noted by Richter, and the steep, outward slope of the cirque's floor in central areas in which rivers were probably still active.[101]

A single series of cirques occurs on mountains which just projected above the snowline, two or more, as in Norway, the Carpathians, Alps and British Isles, where the mountains were severely glaciated and the snowline was much depressed.[102] Reconstructed glaciers may have steepened and recessed the ower cirques in the Lake District.[103]

Aspect. The orientation of cirques furnishes further proof of their glacial origin. In the northern hemisphere, it is mainly northerly or easterly (the remaining slopes often have unaltered funnel-shaped gullies). The

asymmetry is due to snow gathering on the shady and wind-shadow slopes[104] where it lingered because of the severe frost and may be influenced by structure[105]—to-day the snowline is probably about 200 m lower on the shadow sides of mountains. Most modern glaciers face northwards,[106] as J. J. Scheuchzer[107] early observed and Enquist[108] emphasised in the Swiss and Austrian Alps: the glaciers on the Italian are smaller than those on the French or Swiss side of the Alps.

This aspect may be seen in almost all glaciated mountains, as in Norway,[109] Vosges,[110] German Mittelgebirge,[111] Pyrenees,[112] Alps,[113] Hohe Tauern,[114] Niedere Tatra,[115] Carpathians,[116] Dinaric Alps,[117] Apennines,[118] the Balkans,[119] Anatolia,[120] British Isles,[121] Iceland,[122] Alaska,[123] Labrador,[124] west Canada,[125] U.S.A.[126] and Korea, Formosa and Japan.[127] In the southern hemisphere, they also face polewards or to the lee, e.g. in Tasmania[128] and the Peruvian Andes.[129] In the tropics, they look westwards, prevailing winds rather than sun's altitude being the determinant.[130]

The orientation is less pronounced in polar lands because the sun's course is circumpolar (wider temperature oscillations on the south side give the cirques this aspect in Severnaya Zemlya[131]). Exceptions are numerous on the highest mountains in the tropics whose glaciers depend less upon favourable conditions for their growth and preservation[132] and on high mountains elsewhere which because of their great height tended to have a symmetrical glaciation.[133] There are exceptions too in both the Old and the New World if the preglacial valleys were better developed on the south and west,[134] if the daily alteration of freezing and thawing was stronger on the southern flank,[135] or if, as in parts of western North America, westerly winds were the moisture-bearers.[136]

Asymmetrical ridges. A closely related feature is the asymmetrical ridge. Northern faces on east–west ridges and eastern faces on north–south ridges are often steeper,[137] the one set being concave, the other smooth, regular and convex: the junction, frequently sharp, has shifted at the expense of the convex side. The causes of the asymmetry are probably various and include longitudinal crevasses along the shadier side,[138] snow-caps and nivation,[139] or excessive debris which checked ice-erosion on the sunny side.[140]

Methods of glacial erosion. Although geologists almost unanimously favour the glacial origin, they are strongly divided upon its manner and degree. Cirques are unquestionably bound up in some way with cirque glaciers. This is proved by ice-filled cirques ("active cirques"[141]), the smooth and ice-worn floor, the rock-basin, the cirque-moraine, and by the coincidence, as in the east Pyrenees, of cirque and glacier.[142] Yet the diametrically opposite views of protection and profound erosion (see ch. IX) reign here; cirque glaciers, it is said, are protective[143] since cirques are essentially preglacial in age and origin. Alternatively, cirques have been eroded vertically, as in Snowdonia,[144] by 200 m or more metres. The difficulty, as mentioned already when dealing with the general question (see ch. IX), is partly because the operation is largely or completely hidden from sight.

Meteoric action. Frost-riven material from the surrounding walls falls upon the ice which transports it to the snout to make the end-moraine. This process, imperfectly recognised by early writers including K. Loranges (1868), was noticed by Helland[145] and convincingly described by Richter[146] in Norway and the Alps, and has since been emphasised for many cirque-regions,

especially in the Old World.[147] It has even been elevated to be the main cause of cirque formation,[148] notably for the Antarctic where the attack is brisker above than beneath glaciers.[149] This meteoric action was helped by avalanching which, working in early glacial time along rock-junctions, has been regarded as the sole force,[150] and by frost acting powerfully along Ampferer's "black-white" boundary[151] and in the *Gehängekare* during early and late stages. It found a strong ally in the snow-cornices which the wind created on the crests[152]: they tore off astonishing amounts of rock when they broke away.[153]

Though this action was responsible for the sharp crest-lines above the cirques, it alone was not able to make the cirque, notwithstanding the undoubted extension of the action below the "black-white" boundary by repeated freezing of the considerable quantities of melt-water which descended far behind the snow and ice, especially if there was a *Randkluft* or *Randspalt*[154] (see below). Like the hypothesis of bergschrund sapping (see below), it is merely an auxiliary process, working peripherally. It did not deepen the cirque or excavate the floor and rock-basin,[155] nor was it capable of eroding those, the largest of all Alpine cirques, that lay completely under the Pleistocene ice (three-quarters of the Karwendel cirques were so situated[156]). In these cases (cf. p. 297), cirque formation was probably at a standstill[157] and meteoric frost operated solely during the earlier and later phases of each glaciation. Hence its importance should not be exaggerated: it was merely one factor.[158]

Direct glacial erosion. Abrasion and plucking have been frequently charged with making cirques.[159] Andrews,[160] in particular, thought that these were initiated under glaciers as steps that receded towards the glacial divide, the erosion increasing down the cirque wall as the ice flowed more rapidly. Some of the Pennine cirques were regarded as gigantic holes scoured by ice eddying round the valley curves[161] like the glacial scoups that sometimes scar U-valleys on the concave sides.[162] Such forms would apparently include the *Durchgangskare* of N. Krebs[163] or Distel's *Durchgangsmulden*.[164]

Abrasion and sapping operated without question on the floor of the cirque where the flow was in general parallel with the rock-face: they gnawed out the rock-basin (see p. 274).

Bergschrund sapping. The significance and importance of the bergschrund (see p. 45) in connexion with cirques were first stressed by W. D. Johnson[165] who was lowered *c.* 45 m into the bergschrund of Mount Lyell Glacier, Sierra Nevada, California. He found that the rocks in its lower part were shattered into loosened blocks and fragments by localised frost, due to fluctuating temperatures and to waters freezing in pores and structural planes, so that rocks, inextricably interpenetrated by ice, were plucked out as the glacier moved away. Similar observations were made in Scandinavia[166] and in the Alps[167] where melting and undercutting of the cliff head were observed. Erosion of the order of 1 cm/diem was found in the transverse crevasses that meet the glacier bed on the east slope of Pelvoux.[168] Similar freezing occurs in the *Randkluft*[169] (see below).

Basal sapping probably extends downwards as well as backwards. But the downward fracturing is hindered by the difficulty of removing the disintegrated material so that the cirque-floor remains approximately flat, enlarging

horizontally in the trail of the receding cliff. This recedes most during the early stages and where the rocks like granite or other crystalline masses are well jointed or the strata are horizontal and of variable hardness. The potency of the sapping may be gauged from the huge blocks to be seen in these positions; in the Antarctic, several up to 6 m long have been extracted and transported 45 m from sockets still intact and visible.[170] Plucking was most severe at the back wall because there the ice was coldest and most rigid, moved forward at the highest angle to the rock-face[171] and had a constant supply of angular tools.[172]

Johnson's view has been adopted generally in America[173] and by a number of geologists elsewhere.[174] It is confirmed by the absence of cirques in those parts of the Alps which during the Glacial period were completely wrapped in ice and had no bergschrund[175] (cf. p. 297) or where the preglacial slopes were gentle so that the rocks were unable to deliver material to a bergschrund.[176] It is confirmed too by the sharp shrundline (*Schrundlinie*) which Gilbert[177] detected in the abandoned cirques of the Sierra Nevadas. This is found in an occasional "live" Alpine cirque[178] and sometimes accompanies a marked terrace: it separates a rounded, glaciated surface below (where the ice was active) from a steep, raw face above (where the snow and ice were frozen on to the rock).

Sapping clearly depends upon the diurnal or, more probably, seasonal or shorter period oscillations across the freezing point. This is made possible either by the communicating air in open crevasses or the downdraught in the bergschrund, and incidentally in other crevasses by waters that drip from the surface or issue as seepage springs of the internal mountain drainage at or near the foot of the cirque-wall,[179] especially in well-jointed and well-bedded rocks (see below).

Bergschrund sapping, like the meteoric hypothesis, is only an incomplete explanation. Cirques cannot thus arise if, as in certain modern glaciers,[180] including the Antarctic, or in the Pleistocene Alps,[181] the bergschrund is absent or invisible (the bergschrund and its sympathetically curved crevasses may be an effect rather than a cause[182]) or if it does not encounter the wall—as some assert is generally the case[183] (contact is only made if the wall slopes at 50°[184])—or, as usually happens, intercepts the wall far above its foot.[185] The objection is also made that the sapping is too hypothetical and rests on too limited an observational basis. Moreover, temperature changes at the bottom of the bergschrund, unless in small cirques, are exceptional and contrary to experience[186] and do not take place in winter when snow fills the crevasse[187]—in tunnels through the Tyrolese glaciers the ice was firmly frozen to the underlying rock[188] and a series of ice-floors (*Böden*) extended from wall to wall, entirely cutting off all air-circulation or percolating waters. They are most probably absent in the cold Arctic[189] and extremely cold Antarctic[190] and in temperate lands where the bergschrund is usually well above the orographic snowline,[191] and are least likely on just those northern faces where cirque glaciers are apt to rest.[192] In any event, it is necessary to distinguish between temperature oscillations in the air and in the ground.[193] The action is improbable, except possibly in dry regions[194]; for the temperature oscillations are too infrequent to have any appreciable effect,[195] unless, as may well happen, the ice, packed against and frozen to the joint-ruptured rock of a valley-head, pulls joint blocks away without the intervention of the freeze-and-thaw process.[196]

In fine, sapping takes place at the base of a bergschrund (which marks the upper limit of glacial erosion in a cirque) in small cirque glaciers and in the maturer stage of development (in large cirques, the bergschrund migrates upwards and only touches the back wall far above its junction with the floor). It is greater in lower latitudes and in shallow, open bergschrunds.[197] Bergschrund sapping is a main cause of the backward recession of the walls.

The point of attack shifts in various ways; the bergschrund opens at a slightly different position each spring; the glacier rises and falls with varying conditions of supply and wastage; and in the final retreat, the bergschrund steadily traverses the whole of the lower part of the cirque wall.

Emphasis is also laid on the action of waters which, as in modern Iceland, Spitsbergen and the Alps, with active melting, drip or pour into the randkluft and bergschrund and along the head wall.[198] This water, which freezes on high mountain peaks and ridges to form the ice-apron, extends the sapping to the base of the bergshrund if freezing occurs. The most important function of these waters may be to transport the debris produced by frost action and cause the soaking to facilitate such action.[199] The sapping is also associated with ground-water seeping out of the rock,[200] the schrundline (see above) indicating the sapping level. Thus in one way or another, the head wall, as in many British cirques, is shattered and plucked to its base and repeatedly meets the abraded floor in a sharp angle.

The cause of the hollowing out of concave floors near the back of the cirque may be connected with this headwall sapping which may work in an inclined direction, i.e. downwards as well as backwards,[201] though the lateral sapping, in general, probably exceeds the downward erosion,[202] e.g. in those cirques which occur in the angle at the margin of monadnocks.

Subglacial frost action. The lowering of the floor with its rock-basin must be sought in other processes than those just mentioned. Among these may be the freezing and melting which accompany the changing pressures of the ice upon its bed,[203] since increased pressure raises the melting point, decreased pressure lowers it (see p. 115). Water in joints or pores in the rock or in the *moraine profonde* (where this exists) alternately freezes and thaws in unison with the changes and shatters the rocks. Experiments[204] reveal that stones embedded in ice and subjected to varying pressures disintegrate mechanically in the same way as when they submit to changes of temperature alone. The action is independent of a glacier's thickness and the actual pressure so long as the sole is practically at melting point, as is the case for example at the base of a cirque-glacier (see p. 107). Frost action has been traced on *Fenster* exposed by the retreating Vernagtferner, on boulders completely entombed in ice (some of them totally disintegrated), on stones that emerged from inner moraines, and in a gallery piercing the Tête-Rousse.[205]

Salomon[206] in particular has stressed the geological significance of this force. In his opinion, it operates at all inequalities and breaks of gradient and attends the constant opening and closing of crevasses at the base. Its effect is severe because the ice continually removes the riven material and, incidentally, provides angular pieces for abrasion downstream. Blocks displaced on recently vacated glacier beds may be due to its action rather than to plucking (cf. p. 250). It probably took place at roches moutonnées where the ice froze on to the lee side[207] and was enhanced if waters, undercooled by increased pressure, reached bedrock.[208]

Yet such mechanical action is denied.[209] Pressure changes of this kind are,

it is said, either absent or inconsiderable and restricted to relatively few places; basal temperatures are always above, and ground temperatures, as in Spitsbergen and north-east Greenland, always below freezing point; and neither striated and polished rocks nor boulders in the *moraine profonde* show any sign of it.

Its quantitative effect is difficult to judge though Hess[210] sought to do this on the Hintereisferner. Experiments[211] suggest that it varies with the type of rock and produces only fine material like sand, gravel and mud. A single freezing gives merely a fine powder.

Fracturing of subjacent rocks by cooling or changes of temperature alone is probably quite negligible,[212] though frost acting on ice-free surfaces at the beginning and end of each glaciation helped to loosen the blocks.[213]

Nivation. In general, sedentary snowfields, stagnant ice-caps and dead glaciers protect the rocks beneath them from frost and river-erosion. Yet snow-erosion or "nivation"[214] is a powerful and rapid agent in initiating and enlarging cirques. The germ of the nivation idea, i.e. that snow-masses keep the rocks moist and erode by repeated freezing and thawing, is to be found in a paper by Helland[215] in 1876. Stationary snow-banks, notably if they melt gradually and allow the waters to soak into the disintegrated rock, steadily deepen any pre-existing depression, however slight. Excessive frost acts vigorously during summer along the receding edges of the snow at the triple contact of snow, rock and air and also, importantly, below the snow-patch[216] where the permafrost is locally near the surface. Melt-waters penetrate the rocks, especially if these are soft and permeable, and freeze at night and remove the finely comminuted material. Such snow-banks, whose limits vary in response to the many changes in precipitation and temperature, make the rocks around the margin loose and porous by repeated freezings and thawings. In a self-stimulating manner, they sink with indefinite boundaries into the face which becomes concave in profile (see fig. 27, p. 89). Scour and transportation are absent, unless the snow by increasing thickness and pressure passes into moving ice.

The reality of nivation is now established.[217] Snows moisten the ground, notably on fairly steep slopes,[218] and aid solifluxion,[219] though Bowman[220] who believes they have motion attributes nivation to nival plucking rather than to frost. De Martonne[221] ascribes to nivation the multiplicity of cirques in mountains like the Carpathians that had only a restricted local glaciation. It is highly important in western North America[222] and in Greenland where W. E. Ekblaw[223] recognised three types of snow formation and action: (*a*) dome-shaped drifts on plateaux which form terraces by melting progressively inwards from the edge; (*b*) piedmont drifts along the foot of extensive cliffs on lee slopes which sapping attacks at the top and solifluxion at the bottom; and (*c*) wedge drifts in gullies near cliff tops where cirques are initiated. Högbom[224] also described embryonic cirques along joint planes: they are the *rasskars* of Spitsbergen and Norway.[225] In Iceland,[226] snow patches fall into three classes; transverse patches, guided in their slope by horizontal outcrops of beds; longitudinal patches, occupying drainage hollows; and circular patches in slight concavities which develop into cirques. Nivation may also open out small V-forms into small U-valleys.[227]

Doppelgräte,[228] which occur for example in the Central Alps, Limestone Alps and in the Tauern, arise if frost, nivation, deflation and possibly water act on rocks of varying hardness and resistance[229] and in connexion with

snow troughs. On east–west ridges, they may create ridge- and saddle-lakes.[230]

Nivation is regarded not only as the cause of asymmetrical crest-lines[231] and an aid in forming cirques[232] but as a means of excavating these where there was no previous hollow[233] (according to R. Lucerna,[234] nivation turned the growth of the cirque towards the shadow side and formed curved cirques in the Niedere Tauern). It has been emphasised by none so strongly as by Bowman[235] who, basing his conclusions on researches in the Peruvian Andes, showed that snow if thin moved more quickly as the gradient rose, and that the thickness necessary for movement decreased with a rise of gradient. Matthes,[236] from observations in the Bighorn Mountains, thought a snowfield sloping at 7° would require to be 125 ft (38 m) thick before flow began; others[237] estimate the minimum at 150–225 ft (45–70 m): lateral moraines or the height of preglacial modification may be used to obtain these amounts. Bowman greatly reduces it (with a slope of 25° the critical thickness would be 15 m) and recognises that the degree of compactness and diurnal changes of temperature are crucial. He relates nivation to cirque erosion by supposing that a snow-bank resting in a hollow in a slope will deepen from nothing at the lower and upper edges to a maximum where the angle of slope is *c.* 20°. Between the gradients of 10° and 30° a hollow will be eroded that will deepen the snow and by augmenting the pressure at its base convert it into firn and finally into glacier-ice. Further deepening is impossible if the *contre-pente* is 5° until the basal snow becomes ice. Bowman believes the varying rates of motion of snow, firn and ice at a valley head are the real cause of basal sapping at the cirque wall, and that the sharp break between steep head wall and flat floor marks the line at which the erosive power increases and increasing thickness transforms the firn into glacier-ice.

Nivation as a powerful force is demonstrated by the recessed slopes at the edges of snow patches,[238] e.g. in the strandflat gullies in Spitsbergen. It acted widely among the ice-centres of the world at the beginning and end of each glacial epoch, as well as in the mountain clusters on the periphery of the snowline, e.g. in central France and the Carpathians. Whether alone it made true cirques may be doubted—it is denied, for example, for Antarctica today.[239] Nivation cirques[240] or forms resembling the *Karembryos* of Krebs,[241] *Karoids* of Sölch,[242] or *Gehängekare* of Fels[243] resulted; for an action which tended rather to widen than to deepen[244] had probably a definite downward limit though this varied with the climatic environment.[245]

T. C. and R. T. Chamberlin[246] imagine that stationary snow is protective and that erosion begins when moving and abrasive snow-ice coheres to the soil, loose rock or subjacent strata. The line of demarcation between moving and stationary masses is a scar, the cirque's embryo. As the process continues, this becomes more and more pronounced and initiates and localises the bergschrund.

The complete series of gradations between nivation dimples and true glacial cirques arises not by continuing the nivation process but by immobile snow passing into flowing glacier-ice. The change from snow patch to glacier and back again to snow patch encourages cirque growth by removing the disintegrated material.[247] Chemical action was an aid.[248]

Modified preglacial features. While, therefore, small cirques may originate through nivation, as on the Norwegian plateau in places unconnected with valleys, the majority inherited hollows in which ice-sculpture succeeded

water-sculpture.[249] This conclusion was delayed by making the cirques wholly dependent upon the snowline (see above). It has now, however, been demonstrated not only for the Alps[250] but for many other regions,[251] including Scotland, the Faeroes, Pyrenees and German Mittelgebirge, which being lower have been less severely modified. Austrian geologists, in particular, have emphasised this view since 1920 by stressing the importance of the preglacial relief and minimising that of the ice. It may indeed be as true to say with Agassiz[252] that glaciers exist because of cirques as it is to state the converse.

The half-funnel shaped *bassin de reception torrentiel* (I. C. Russell's "alcove"[253]) has been converted into a half-cauldron cirque. By adding its own "parasitic relief",[254] the ice has modified volcanic craters,[255] niches in mountain sides,[256] landslip scars or fault hollows,[257] spring amphitheatres and karst depressions,[258] ravines (especially those, as Richter emphasised) which are radially grouped, and the beginnings of hydrographic nets,[259] either along the sides or at the head of main drainage lines—the cirque floors belong to the region of the trough shoulder and unite into a single surface.[260] The slope has been steepened into a precipice; the base has been thrust back; and the floor has been flattened or inclined backwards.

All stages in this transformation of ravine into cirque have been recognised by Richter[261] and traced in the field as in the Alps,[262] Pyrenees[263] and South Victoria Land.[264] Davis[265] deduced them from studies in Snowdonia and central France. Doubts may exist about the category to which a particular hollow belongs[266]: many rain-funnels in the Zillertal, for instance, are locally called *Kare*.[267]

The steeply falling floors of many cirques, as Richter, Salomon and de Martonne noticed, and the presence of such cirques in more highly domed parts of the *Gipfelflur* (see p. 327), e.g. the Zillertal Alps,[268] suggest this derivation. Further pointers are the steps which may mark ruptures in the preglacial slopes[269] (see p. 297) and the threshold's probable coincidence with the confluence of preglacial streams[270] or constrictions in their courses.[271] More conclusive are the cirque's disposition respecting the present drainage; its position next to unaltered stream funnels,[272] as in the Riesengebirge and east Carpathians; the river-eroded hollows in southern and western slopes[273] (see p. 298); the actual head of a preglacial valley occasionally preserved above the cirque[274]; the passage from U-shaped cirque above the end of a glacier to unaltered V-shaped ravine below[275] (cf. p. 322); and winding cirques as in the Karwendelgebirge.[276]

Salomon,[277] discussing the evolution, distinguished a cirque-embryo or "Richter stage", with mainly backward weathering of the walls, and a later "Mösele stage" (from Mösele, Zillertal Alps), characterised by powerful ice-erosion of the floor when the snowline was further lowered (H. Philipp[278] thought this too was an early stage and not a development of the cirque embryo).

Cirques commonly coincide with the head of a rejuvenation,[279] as has been frequently demonstrated in the eastern Alps and Tyrol.[280] The altitude of the floors in the Karwendelgebirge was controlled by the valley system and preglacial relief and was a function of the length of the tributary.[281] Cirque and valley floors fall into the same zone with a direct and uninterrupted transition from *Hochflur* to cirque, e.g. in Rondane and Jotunheim.[282] In areas of low preglacial relief, cirques are unusually flat and have poor thresholds

and no *Rücktiefung*.[283] Twin cirques dwell in preglacial valleys which forked at the upper end, and more complex forms where many valleys existed. The absence of typical cirques from the highest Karakoram and Hindukush mountains is due to the topographical uniformity,[284] and from the western Alps and central Caucasus to the steep fall.[285] The restriction of well-developed cirques in the central Alps to the Gotthard region[286] may spring from the *Hochgebirgs* nature of the preglacial relief[287] or the rapid growth of the firn-cover[288] which the great upheaval of this part of the Alps, postulated by H. v. Stapf on other grounds (see p. 327), has caused.

Cirque recession. Meteoric action above and sapping and plucking behind compel cirque-glaciers to cut back abruptly. This "head wall recession", whose seriousness in glacial sculpturing Helland and Richter were the first to appreciate,[289] scallops the mountain with a "biscuit cutting effect"[290] if, as in the Cuillin Hills of Skye,[291] the cirques are numerous enough—the relatively long cirques of Rondane in west Norway are due to the ease with which the local sparagmite was cut back.[292] For full development, broad massive mountains are essential[293]—Richter,[294] from Alpine observations, judged that their base should be three times their altitude and the angle of slope should not exceed 31°. "Crater cirques"[295] arise in sub-summit positions as in coastal Norway north of the Arctic Circle.

At an appropriate stage in the glacial cycle, the importance of cirque encroachment and recession is paramount, exceeding that of valley glaciers. Hobbs,[296] who realised that the work of cirque-glaciers was a function of their duration, distinguished several stages in the topography's progressive evolution (fig. 58). In his Grooved or Channelled Upland, Penck's *Rundling*,[297] the broad massive ridges and dome-like forms, such as Mont Blanc and the *Kopf* of many mountains in German-speaking countries, are little dissected and the preglacial relief is largely preserved. After the cirques have eaten back strongly and several of them, by convergent retrogression, have consumed a central peak the *Karling* results (typical of the Cuillin Hills[298]), the cirques communicating with each other by "*Karling* passes"[299] which have saddle-shaped or hyperbolic curves. Further encroachment and sharpening of the ridges produces the Fretted Upland, well seen in the Alps, Lofoten Islands, the High Sierra Nevadas and Cascade Range of North America, and in the Royal Society Range, Antarctica. In this type, the preglacial relief is almost wholly dissected (the *Firnfeldniveau* in the Oetztal Alps is in places completely destroyed[300]) and replaced by a meandering skeleton comb-divide and a network of arêtes, main and lateral, with sharp peaks and horns and cusp-shaped passes, the Törl type of Sölch's classification.[301] In Hobbs' Monumented Upland, exemplified in the Glacier National Park, the recession has left only isolated peaks; the thin arêtes have been eliminated and superseded by meandering divides and unusually low cols, the Alpine *Gletscherjoch*. In the Cascade and Rocky Mountains of western America, such passes have been followed, first by Indian trails, then by highways and railways.[302] O. Maull[303] recognised a series beginning with *Quelltrichter*, *Talschlüsse*, *Hochtäler*, *Schluchten* and *Wandnischen* and evolving through the *Wannenkar* into the *Grosskar* and finally into the *Karterrasse*.

The various stages may be exhibited in a single group, as in Jotunheim,[304] and in Arran[305] where the slopes are maturer in the east. They are also arranged one above the other, as in the Alps,[306] *Rundlings* occupying the

lowest, *Halbkarlings* (cirques on one side only) the next, and *Karlings* the highest positions.

The recession of inosculating cirques creates zig-zag knife edges, with *dents* and *aiguilles*, so well described by J. Ruskin.[307] They are the culmination of

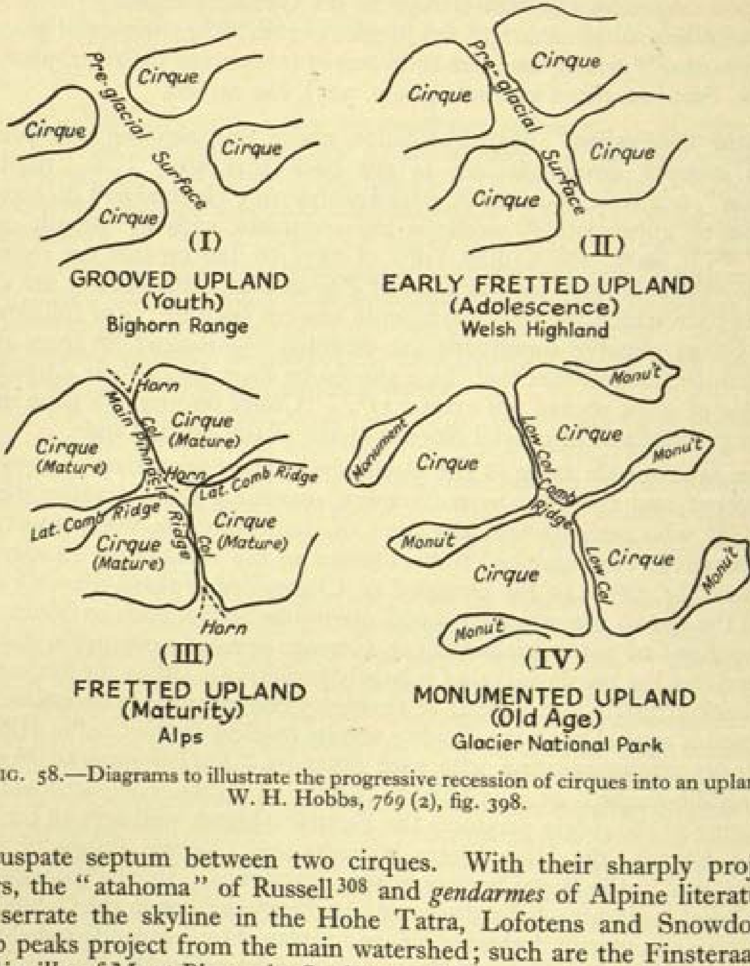

FIG. 58.—Diagrams to illustrate the progressive recession of cirques into an upland. W. H. Hobbs, *769* (2), fig. 398.

the cuspate septum between two cirques. With their sharply projecting towers, the "atahoma" of Russell[308] and *gendarmes* of Alpine literature,[309] they serrate the skyline in the Hohe Tatra, Lofotens and Snowdonia.[310] Sharp peaks project from the main watershed; such are the Finsteraarhorn, the *Aiguilles* of Mont Blanc, the Snehätten of Norway, the Cobbler of Argyll, and other peaks in Arran, Mull and Skye.

If three cirques come together, the mass of rock sweeps up with incurving sides to a tricuspate peak; where four cirques meet, to a quadricuspate peak, as the Matterhorn; quinquecuspate peaks like the Mönch are rare. These horns are widespread, e.g. the Aletschhorn, Wetterhorn, Ritzlihorn, Breithorn, Nesthorn, Weisshorn, Shreckhorn, Gross Glockner, Titlis, Jungfrau, and L'Aiguille du Dru in the Alps, Mount Sir Donald of the Selkirks, Mount Assiniboine in the Canadian Rockies, and others in South Victoria Land and the King Oscar Coast, Antarctica.[311]

Hence, cirque recession consumes mountains and converts *Mittelgebirge* with domed watersheds into *Hochgebirge* in which cirque and "alpine" scenery predominate.[312] Since the latter, being linked with the glacial

snowline, may occur at any altitude from above 1500 m in central Europe[313] to practically sea-level in the Lofoten Islands and arctic regions (see p. 296), and rounded forms are to be observed in many parts of the high American Rocky Mountains, the term is somewhat misleading.[314] There are transitions in the Transylvanian Alps.[315]

Cirque-terrace. It has been conjectured that glacier troughs or U-valleys of considerable dimensions may originate by headwall erosion of cirques.[316] It is also thought that by cirque recession and the lowering and removal of partition walls and their towers, cirque floors may be enlarged and united into a new denudation level, the "cirque platform", "cirque-floor bench", or cirque-terrace[317] (*Karplatte*[318]). Richter[319] was of the opinion that wide plains, such as the Norwegian palaeic surface (see p. 348), resulted from this "decapitation of mountains". Others attribute to it the shoulders of U-valleys,[320] the *Firnfeldniveau*[321] (see p. 326), the Norwegian and Spitsbergen strandflat,[322] and the low coastal platforms of South Georgia and Graham Land.[323] Lucerna[324] has even (erroneously[325]) derived *Mittelgebirge* from *Hochgebirge* in this way.

Such terrace-formation corresponds more to a deductive scheme than to actual observation,[326] except as relatively narrow strips of irregular relief and considerable slope. Glaciation was not long enough to permit this final stage to develop. Moreover, as the recession approaches completion and narrows the gathering grounds, it automatically forces the glaciers to dwindle, destroys the conditions upon which they depend, and brings about their own extinction[327] (see p. 213)—Haast,[328] for example, thought the ridge-forming action contributed to the retreat in New Zealand.

Date of excavation. Glacialists differ widely about the age of the cirques. These are referred to the early stages[329] of each glacial sub-cycle, when frost was severe, their moulding and rounding belonging to the maximum of the regional glaciation which striated their backs, made them less steep and erased the pinnacles of the fretted arêtes. The early partition walls on the Scandinavian divide were in this way removed where they stood across the flow from the later, eccentric iceshed[330]—the partially erased cirques are named *kjedels*—or have kame-terraces or other marginal features.[331] Tyrolese cirques were also overridden transversely or obliquely,[332] as were the earlier cirques in Labrador,[333] Greenland[334] and north Finland.[335]

Taylor's "palimpsest theory"[336] supports this view. It supposes that the cirques of South Victoria Land were couloirs which cirque-glaciers of Pliocene or early glacial times modified and later plateau ice from the west overrode and severely altered, leaving relics in the shape of cirque-basins, rock-barriers and terraces. The cirques in Ross Quadrant,[337] which are so low as to be partially or completely submerged by Ross Barrier or by ice from the larger glaciers (similar ones, older than the general glaciation, occur in South Georgia[338] and may lie beneath the steeper falls as in Ferrar Glacier[339]), have been regarded as a strong argument for this early date.[340] While such *Durchgangskare*, as Krebs[341] termed them, may have so originated, this can not be true of the *Ursprungskare*.

Many of the big cirques in Petermann Range and other mountains of north-east Greenland, where the interval since the complete inundation of the area by ice was relatively short, were apparently also initiated and developed in part during the advancing hemicycle,[342] and like those of the Torngat

Mountains of Labrador[343] were spared destruction, though they were later overridden and toned down by the continental ice. Some are being uncovered by the present recession.[344]

A lateglacial date has been given for some mountain groups south of the North American ice-sheet[345] (White Mountains, Adirondacks, Catskill Mountains, Mount Katahdin), Scottish Highlands,[346] Norway[347] and parts of the Alps.[348] Lucerna[349] has frequently advocated it in extending to cirques his ideas on the excavation of U-valleys by the ice of the "postglacial stages" (see p. 321). But it is denied[350] because interglacial breccias, as in the Karwendelgebirge,[351] are still intact, and the lateglacial moraines and the cirques behind them are markedly disparate in size. This association of diminutive moraine and large cirque is almost universal, recorded exceptions being rare.[352] While there may have been severe loss into glacier-streams,[353] because the glaciers thoroughly comminuted the eroded rocks (see p. 219) or the debris, as in the White Mountains,[354] was spread on to extraneous ice encircling the mountains, in almost all cases it must be assumed that the sculpturing and removal of the material belonged to a time when the glaciers, whether lateglacial or not, were still sufficiently big to carry their loads far away from their source. Interglacial subaerial erosion may also have played its part.[355]

Yet a third view places the excavation at the climax of glaciation[356] when the debris was carried far afield. The cirques were fashioned by activities now practically inoperative.[357] The short "postglacial stages" of rising snowline left the *Rundlings* unaltered except for the cirque-basins and the steepened floors.[358] But this date is dismissed by supporters of the meteoric and bergschrund hypotheses who think the cirque recession was then at a standstill.[359] It was only possible if pronounced relief, as in the English Lake District,[360] favoured flow from the centre. Yet there can be little doubt that, as in the British Isles[361] and Labrador,[362] some cirques are early glacial, others lateglacial, *hochglacial* forms occurring only if the ice did not submerge the whole.

Conclusions. The contrast between raw walls and glaciated floors bears witness to different forces in the birth of cirques. The majority have developed from pre-existing hollows by (*a*) meteoric action above the ice, (*b*) bergschrund sapping and plucking of the walls, (*c*) freezing of melt-waters at the head wall below the bergschrund—this action is probably slight—and (*d*) plucking helped by frost resulting from changes of pressure. The floor was abraded by ice, shod with angular material constantly replenished from falls into the bergshrund, and from sapping and plucking below and behind. This detritus was carried, according to Finsterwalder's theory (see p. 119), along the lower layers of the glacier. Plucking and abrasion were responsible for the cirque-basin (see p. 274).

Some of the excavation dates from early glacial time and from maximum glaciation in those cases where the ice did not completely bury the hollow. Much is referable to the lateglacial retreat when nivation was severe and a rising snowline left its stadial cirques on the mountain sides. The cirques were re-occupied and refurbished with each glaciation.

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Chapter XIV

U-VALLEYS

Form. By comparing unglaciated with glaciated valleys, e.g. valleys in the southern and eastern Alps with those in the rest of the Alps, including the Zillertal Alps and other domed parts of the *Gipfelflur*,[1] certain distinguishing morphological features became apparent. The V-shaped cross-section of the river-valleys (strictly this only applies to steep, youthful valleys, the stable form in mature valleys being the catenary curve) is replaced in glaciated valleys by a round-bottom V or by the U-shape which, noted by Helland[2] in 1877 and later by McGee[3] ("glacial canyon"), gives them their generic name.

U-valleys vary noticeably in their shape according to the kind of rock. The typical U or "parabolic valley"[4] characterises hard and homogeneous rocks,[5] as in the crystalline Alps or Tyrolese Muschelkalk, especially if the preglacial precursor was sharply defined[6] (see p. 321), but grades into a graceful concave curve[7] in soft strata like slate or schist which preglacially had probably attained broader forms. Screes and avalanche cones on the sides and aggraded floors mask the true form; they may convert it into a trapezoid[8] (Ger. *Kastentröge*). Steep valleys may be convex and V-shaped[9] (see p. 323).

The cross-section often changes downstream. Here the preglacial streams eroded laterally and the Pleistocene glaciers flowed more rapidly,[10] expanding on to the foreland to make the valley trumpet-shaped.[11]

Lateral valleys are often markedly discordant as A. Schlagintweit[12] noticed in the Alps in 1850 and J. Hector in New Zealand in 1863. Streams, sometimes harnessed for power, dash over the walls in a series of short cascades (*Giessbach*, *Pissevache* or *Salto*), half hidden and only triflingly incised. Magnificent examples of these *gradins de confluence* (Ger. *Stufenmündungen*[13]) occur in Lauterbrunnen (e.g. Staubach) in Switzerland and in the Yosemite Valley, California. Ascent from the main valley is often difficult (see p. 499). The Brenner valley only attains the valley of the Sill at Matrei, and the Tauern railway climbs about 8 km from Schwarzbach along the hillsides to reach the Gastein Ach above its step.

These perched or "hanging valleys" (Ger. *Hängetal*; Fr. *vallée suspendue*), as Gilbert[14] termed them, often possess a barrier at their mouth[15] (Ger. *Mündungsriegel*). They have a well-developed U-cross-section, as in the Zillertal and Oetztal Alps,[16] and pass through the transition of short U-shaped hanging valleys into cirques.[17] Subtypes, based upon the degree of modification, have been distinguished.[18]

Pronounced salients of solid rock which project from a few trough walls below or at the lips of hanging valleys are termed bastions[19] (Ger. *Felsvorbauten*; *Stufenvorbauten*): a fine example occurs at the mouth of the Trient valley. Their presence indicates a weakening of the erosive power of the trunk glacier as a result of thrust exerted at right angles to it from the tributaries,[20] of a reduction in the rate of flow due to the accumulation of debris at the confluence of the glaciers[21] or at discordant junctions.

As Rütimeyer[22] observed, shoulders, benches or shelves, as they are variously called (Fr. *épaulement*), slope backwards and upwards from the

Trogrand at the top of the "lateral scarps".[23] Their steeper part grades into the flat, high mountain meadows of the Alps which carry a variety of local generic names,[24] such as the "alp" which gives its name to the range (e.g. *Wengernalp* and *Grütschalp* on either side of the Lauterbrunnen trough), *piani* in the Italian, *replats* or *replatons* in the French, and *Ebenen* in the German Alps. They sometimes pass into the cirque-floor or cirque-terrace (see p. 307), though these generally have a lower angle. The shoulders may be roughly constant in height or, as in the French Alps,[25] irregular. Often rounded and striated, they bear lateral moraines and wastage, and are rarely absent though they are said to be wanting in the Limestone Alps[26] and the Caucasus.[27] They may be single, double or multiple as is implied in their popular designations, e.g. the prefixes *ober*, *mittel* and *nieder* of the German Alps.

The floor of the U-valley is rarely smoothly graded: in such cases, as in the great eastern valleys of the New Zealand Alps,[28] the grading is of postglacial date. The longitudinal profile almost invariably reveals a step and tread arrangement (*le profile longitudinal en escalier*). The cyclopean treads of the rude stairway (Ger. *Taltreppe*[29]), especially in the higher valleys[30] (*Kartreppe*,[31] see p. 297), are crowned by rock-barriers and gently reversed to form rock-basins, e.g. the "paternoster lakes"[32] in the upper reaches and moraine-dammed lakes below. The steps, sometimes hundreds of feet or metres high, cross the valley from side to side roughly at the height of the steps in the tributaries. The uppermost is the trough head, the *Talschluss* or *Trogschluss* of Swiss geologists (in some Alpine valleys present glaciers conceal it), where the two walls of the trough come together. Above it is the *Trogplatte*[33] which grades upward into the floor of a large terminal cirque or number of cirques. Exceptionally, the *Talschluss* (Fr. *vallée aveugle*; "trough-end" or "cross wall"[34]) may be missing[35] or replaced by a flight of steps.[36]

A notch sometimes extends above the shoulders about the height of the *Schliffgrenze* (see p. 40). It marks the glacier's upper limit[37] or the biggest of its retreat stages.[38] This *Schliffkehle*,[39] with its bounding *Schliffbord*, may be a striking feature along the side of a nunatak or mountain (pl. XIA, p. 352) especially near the Talschluss or at the erosion scoups (Ger. *Prallstellen*) of the glaciers[40] where the hillside is sufficiently steep or is interrupted by a preglacial feature. It is generally poor or missing, as where hanging and trunk glaciers were confluent or below the level of the snowline or where subsequent frost action was severe. It inclines outwards and intersects the terraces made by differential weathering. It may have been eroded by ice[41] or more probably by frost[42] or plucking[43] along the flank of a glacier or the randkluft in the firn basin.[44] It may modify a pre-existing feature,[45] e.g. middle Tertiary surface, or notch a previously continuous slope.[46]

Non-tectonic. After the diluvial theory, advocated among others by G. Buffon, G. A. Werner and P. G. Pallas, had been relinquished, early writers like A. v. Humboldt, L. v. Buch, É. de Beaumont and W. Hopkins, regarded valleys as tectonic (cf. general history of the question[47]). This view was held in Switzerland until after the middle of the 19th century, as by E. Desor[48] (1860), B. Studer[49] (1863) and K. Sonklar[50] (1873), although N. Desmarest had clearly reasoned in 1774 that the valleys in central France had been carved by the streams that flowed through them. This epoch-making theory, anticipated too by J. G. Sulzer (1762), J. E. Guettard (1770)

and J. L. Heim (1791) and accepted by H. B. de Saussure (1779) and J. P. B. Lamarck (1802), was later elaborated by J. Hutton[51] and J. Playfair.[52] Nevertheless, it was neglected for many years while the tectonic opinion continued to be held, as by R. I. Murchison[53] for the Alps and the brothers Schlagintweit[54] for the Alps and Himalayas. Re-affirmation of the uniformitarian view[55] by Lyell, Dana and Rütimeyer caused its rival to be abandoned.

The contest is no longer between tectonic forces and erosion, though the former have not been without effect,[56] but between two erosive forces, namely, running water and moving ice. Opinion ranges from a complete denial of ice-erosion, glaciation being equivalent to a relative cessation of valley formation (see p. 212), to the belief that U-valleys were wholly excavated by glaciers during the glacial epochs (see p. 319) or even the postglacial stages (see p. 321).

The problem is beset with difficulties: they include the efficiency of ice as an erosive agent; the action of subglacial streams; the length and number of the interglacial epochs when running water operated alone; and the state of the preglacial relief the Ice Age inherited.

Proofs of glacial erosion. Aside from the faulty argument that valleys were occupied and therefore eroded by glaciers[57] and a supposed necessity for ice-erosion,[58] evidence abounds that valleys were drastically modified as the glaciers passed through them. The glacial *Formenschatz*, features just described as characterising U-valleys, collectively diagnose glacial erosion.

A glacier like any fluid in motion continues to erode until the cross-section of flow corresponds to its own requirements; it fashions its bed to the semicircular shape, Richter's "hemispherical groove",[59] in which the mean hydraulic depth is greatest.[60] The round-bottom V and broadly open U correspond respectively to young and mature stages of glacial erosion,[61] the catenary curve with smooth sides and rounded floor being possibly an even later stage.[62] These maturer forms are well seen in the frontal parts of cirques, in the mouths of many hanging valleys and perhaps best of all in the troughs of former outlet glaciers, e.g. between the Syv Søstre on the coast of Norway. Floods during the melting of the ice may have helped to form the broad floors.

The trough, Richter's word[63]—hence termed Richter's trough[64] (Ger. *Trug*; Fr. *auge glaciare*)—owes its U-shape to truncation of the ends of the forward reaching and interlocking spurs, as noticed[65] early in New Zealand, Kerguelen and North America. Normal river valleys have ravined and buttressed sides and spurs which overlap in rhythmical alternation and trail down from the tributaries. These abrupt bends and turns were unsuited to the ice. They were greatly worn or hollowed out, especially on the impact sides in winding valleys, as is seen on the flanks of modern glaciers,[66] below Sargans in the Rhine valley,[67] and in the nearly horizontal downstream scour on the sides of spurs and residual knobs.[68] The rigid ice, attempting to remove the sinuosities and make a valley only slightly curved, slowly ground away the salient spurs, abrading on the upstream and plucking on the downstream side, and truncating the spurs into a row of blunted gables or craggy, triangular facets. All stages may be traced from the severance of a toe, and its reduction to smoothly domed "beehive" forms[69] or semi-detached knobs ("knob-field") or roches moutonnées on the floor, to its final elimination[70]; the side ravines are then destroyed and spurless walls, unindented by embay-

ments, are created.[71] By aligning its spur-end facets (Fr. *facettes d'éperons tranqués*), the valley acquires straight, parallel sides or curves of large radius as noticed early.[72] Occasionally, however, as in Alaska and in the granites of the Californian Sierra Nevadas, the spurs remain intact, the valleys are V-shaped and still closely follow the preglacial windings.[73]

The hanging of branch valleys, the significance of which was first pointed out[74] for the Milford Sound district of New Zealand (1864) and for the Tatra (1885), is possibly the supreme witness of the magnitude of glacial erosion[75] (cf. below), provided it be remembered that small hanging valleys, perched high above the main floors, may have developed from high-set preglacial niches by nivation or avalanching or from glacial diffluence, and so have no direct bearing on this problem.[76] The hanging reflects the lowering in the main valley and the relative cession of erosion in the tributaries. The Val di Lares, 700 m above the Val di Genova, is probably Europe's greatest discordance.

While tributaries in a river basin normally enter at grade in agreement with Playfair's Law,[77] they sometimes hang if for any reason they have failed to keep pace in their downcutting. Instances have been noted in limestone and tropical countries,[78] as in Columbia River canyons; in youthful faulted areas[79] like Colorado, Black Forest, Odenwald, Jordan Valley, the Rhine near St. Gall and the Indus; in areas where the master stream was favoured by tilting,[80] by rejuvenation[81] (during an important but short-lived stage), by weak structures due to brecciation along shatter belts or outcrops of soft rocks,[82] as in New Jersey and Connecticut, the Alps and the Skiddaw Slates of the Lake District whose hanging valleys lie in the harder Borrowdale volcanics, e.g. south-west and south of Buttermere. Hanging was favoured too by melt-waters issuing from the ice[83], as along the Ohio and Missouri, and by tributaries freezing in winter because they were higher and less copious.[84] It also arose if a main stream undercut its meander scar,[85] from river capture and diversion,[86] and from aggradation of the tributaries,[87] just as the U-form and its truncated spurs resulted in limestone terrain[88] or from non-glacial agencies like faulting.[89]

Although hanging may therefore have developed in a legion of ways, showing the need for caution, the occurrences themselves are quite exceptional. It remains one of the most convincing proofs of glacial erosion; the assertion that hanging, together with U-valleys and overdeepening, is wanting in the glaciated Caucasus[90] is apparently untrue.[91]

Rock-basins in the floors of U-valleys likewise attest ice-erosion (see p. 275): they may be its most decisive proof[92] (cf. above).

The restriction of the U-valley to glaciated terrain is also strongly presumptive. There is none, for example, in the Driftless Area of Wisconsin (see p. 227), where the valleys are narrow and ramifying, or in the unglaciated areas of the south Alps, including the Oglio valley which has accordant tributaries.[93] Corroborative are the contrasted shapes of the valleys in the North Island (unglaciated) and South Island (glaciated) of New Zealand[94]; the definite undercutting at the edge of some modern glaciers[95]; stadial moraines in the Alps,[96] e.g. the sharper relief within the Gschnitz moraines (see p. 1159); the complete proportion between ice-mass and excavation of the valleys in the southern Alps[97]; and the cirques occupied during the last glacial epoch.[98]

Equally suggestive are the deductions of Davis[99] who in his physiographic

study of north Wales deduced the consequences of glaciation, on the assumption that glaciers do erode, and compared them with observed facts. His analysis showed that they are consistent with great erosive power could this be proved.

Deep borings and seismic soundings in the Hintereisferner[100] establish that its bed is trough-shaped and changes abruptly at the sides. The bed is lower than the ideal prolongation of the flanks of the glacier would suggest. Overdeepening has been revealed by the retreat at the junction of this glacier with the Hochjochferner.[101] These observations have been verified on the Floitenkees, Zillertal Alps,[102] and on the Pasterze, whose cross-section is trapezoid,[103] and from determinations of the ice-flow on the Rakhiot Glacier, Nanga Parbat.[104] The same form on the Argentière Glacier of Mont Blanc[105] may be inherited from the Glacial period.[106]

The U-valley is in a coherent view the analogue not of the river valley but

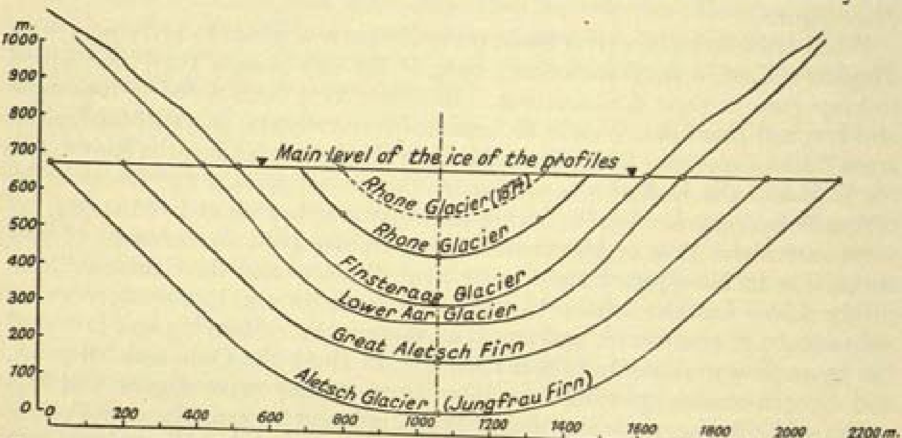

FIG. 59.—Transverse profiles of the Aletsch, Lower Aar, Finsteraar and Rhône glaciers. R. Koechlin, *Alpen* 19, 1943, p. 104, fig. 3.

of the river bed.[107] Pools, bars and rapids, the gradational irregularities of the river, are represented by rock-basins, barriers and steps in the U-valley. They are alike proportioned to the moving medium which filled the channel. A river-bed is U-shaped like a glaciated valley and diminishes towards its source; its shoulder-like banks are the surface counterparts of the shoulders of the U-valley and would be more regular were there no eddies or fluctuating water-levels. Glaciated walls bear traces of nearly horizontal downstream scour analogous to that on the side of stream channels.[108] Erosion scars (Ger. *Prallstellen*) occur on the outside of bends or where tributaries impinged. In other words, both glaciers (at the height of the "ice-flood") and rivers have graded surfaces which obey Playfair's Law and discordant floors that conform to the law of adjusted cross-sections. While in most rivers the volume is greatest at the mouth, in glaciers the cross-section diminishes from about the snowline (in single glaciers) or just below the snowline (in composite glaciers) towards the snout. The parabolic curve of the glacier-bed, as illustrated by the Aletsch, Lower Aar and Finsteraar glaciers and by the recently abandoned bed of the Rhône Glacier (fig. 59), has in general the same form, though this tends to become flatter over its middle stretch in the case of longer and thicker glaciers where the lower ice was more plastic and less erosive.[109]

Wholly ice-eroded? Tyndall's observations, made mainly in the vicinity of Monte Rosa, led him to speculate on the excavation of Alpine U-valleys by prolonged glacial erosion.[110] In this he was supported by G. Bischof[111] and afterwards by numerous workers in all parts of the world,[112] e.g. Alps, Scandinavia, Greenland and North America. The amount of "over-deepening" (Ger. *Übertiefung*; Fr. *surcreusement*) or "underdeepening" (Penck[113] introduced the one and Ampferer[114] the other term) that these geologists demanded is considerable; it is *c.* 180 m in Wales,[115] many hundreds of metres in the North American Cordilleras and Alaska,[116] and 800–1000 m in the Altai.[117] Representative Alpine figures[118] are 270 m in the Traun, 300 m in the Inn and 300–400 m in the Reuss.

These estimates were made by reconstructing the preglacial floor (*Hochtalboden*[119]) in one of five ways, all complementary; (1) by extending the profiles of the hanging valleys over the main valley[120]—the shorter and younger the tributaries, the larger is the error; (2) by lengthening the line of the shoulder into the trunk valley[121] or, if only one side exists (as is often the case), by continuing the curve over the axis of the valley—this method is less good because the altitude of the shoulders is uncertain and may vary by 25 m[122]; (3) by prolonging the *Trogplatte* and the floors above the steps,[123] a reconstruction which gives an extremely wide floor[124] and a more gentle fall than in the modern valley[125]; (4) by continuing the curve of the spurs[126]; or (5) by drawing curves tangential to the rock-barrier.[127]

The discordance of the tributaries varies in one and the same main valley and generally increases towards its mouth. Extreme glacialists correlate this with the size of the tributary glaciers, the discordance being inversely proportional to their dimensions, i.e. small valleys enter over high, large valleys over low steps.[128] If the confluent glaciers are equal, the floors are at grade[129]—even these may have steps due to confluence[130]—and if, as occasionally happened,[131] e.g. in the Durance valley, Pustertal and Hohe Tauern, the glacier in the main valley was smaller than that in the tributary, the main valley hangs. Coefficients for determining the size of a valley and its glacier are admittedly difficult to obtain, while departures from Penck's law are ascribed to rock texture and structure and to currents in the main glacier.

The shoulders are on this hypothesis variously explained. Ice, it is averred, works differently above the shoulders and in the trough, namely surfacewise over the one and vertically in the other with a maximum along the axis where the ice was deepest.[132] But this is erroneous since a glacier's cross-section expands uninterruptedly from the edge to the central line and its erosion should likewise grow steadily and gradually. A glacier does not possess a critical depth of erosion[133] as was imagined by A. Cozzaglio (1895) who, among the first to express the view that ice overdeepened the Alpine valleys,[134] attributed the sudden overdeepening at the shoulder to the critical depth, namely 400 m, at which by basal pressure-melting erosion suddenly set in. Philipp[135] ascribed the downcutting of the trough to its own glacier and the lowering of the shoulder to ice (*Flankeneis*) issuing from the lateral cirques. The curves above the trough may have been lowered 30–40 m.[136] Others suppose the trough was initiated during the first glaciation and accentuated during the succeeding glaciations[137] and was occupied, either early glacially[138] or lateglacially[139] or during the several glaciations,[140] up to the top of the walls only. But it has been repeatedly observed, e.g. in the Adamello group,[141] that the ice rose above the limits of the trough, which in the Alps[142]

was closely connected with maximum glaciation. An ice-sheet may equally well erode U-valleys since along such lines its depth and flow would be greater.

This explanation is not reconcilable with the *Talschluss*. The abrupt descent from the shoulders over the lateral scarp has its counterpart, as mentioned already, in that from the *Trogplatte* over the *Talschluss*. This feature, so intimately bound up with the U-valley's origin, is a complete enigma on this version of the ice-erosion theory which regards the *Talschluss* as a confluence step[143] that arose where glaciers from a cluster of converging cirques heightened the erosion. But there is often no convergence[144] nor any terrace flanking the trunk valley below the confluence.[145] While low steps may have so originated,[146] the *Talschluss*, as is now perhaps generally recognised, had a rejuvenated valley head as its preglacial precursor (see p. 328).

Multiple-trough hypothesis. Rütimeyer[147] observed rock-terraces in Alpine valleys which he referred to separate glacial epochs but others[148] related to separate base-lines. Subsequently, Brückner[149] thought he could trace an interglacial (Mindel-Riss) floor in the Alps, e.g. in the Reuss, Arve, Rhône and Linth, besides a preglacial floor and one 300 m lower. He noticed further that hanging valleys were often double. While Penck,[150] who found a like arrangement in the Italian Alps, correlated the floors with the Pliocene and preglacial, other geologists[151] have accepted Brückner's interpretation: as many as three interglacial floors have been recognised in Switzerland.[152] The alp and the trough have been referred to two cycles of glacial erosion separated by an interglacial period.[153]

Hess,[154] developing Brückner's view to its extreme conclusion, distinguished four floors, approximately equally spaced and belonging to the four Alpine glaciations. He maintained that the snowline rose 200 m at each glaciation; the preglacial surface lay at the *Schliffgrenze*, as Richter early suggested but Penck[155] denied; the glaciers, though 300–500 m thick during each glaciation, became progressively narrower; the U-valley's cross-section was composed of intersecting curves, concave upwards, each scallop steeper than the one above it; the benches corresponded with the curves of the hanging valleys which are similarly composite and discontinuous; the repeated shoulders and steps in the tributaries belonged to as many, usually four, recurrences of glacial epochs; the rivers in the mild interglacial epochs cut V-shaped gorges (*gorges de raccordement*) into the broad glaciated floors and the succeeding glaciations deepened them and rounded their edges, thereby facilitating overdeepening (Fr. *emboîtements des auges*); and that the lower scallops date from the postglacial stages.

Hess extended his studies of profiles from the Oetztal into the Rhine valley and its tributaries, into the Zillertal Alps, upper and lower Engadine, Norway and the Caucasus. His view of the origin of the multiple benched profile (Fr. *épaulements multiples*) has been widely adopted,[156] for example, for the Alps,[157] east Carpathians,[158] Caucasus,[159] Pamirs,[160] Turkestan,[161] Altai,[162] Himalayas,[163] North America,[164] Greenland[165] and east Africa.[166] Each trough has sometimes its moraine and *Schliffgrenze*[167]; gorges below the hanging valleys, as in the Adamello group,[168] show a sequence of stages corresponding to the interglacial epochs and postglacial stages; and, the lowering, calculated from the depth of the valleys and the lengths of the glacial epochs, agrees with the present rate of erosion by the Tyrolese glaciers—the downward erosion by the Günz glaciation was 200 m[169] and

by the Würm glaciation 300 m.[170] F. Machatschek[171] gave the maximum values for each glacial epoch as *c.* 100 m and stated that in the Rhine valley near Chur the total lowering was 1000 m, of which 600 m was interglacial and *c.* 400 m the result of the four glaciations.

The weakening of the forces or, alternatively, the greater smoothness and diminished friction presented to the ice were responsible for the successive narrowing of the troughs.[172]

Although the cyclical hypothesis is perhaps but a logical development of the glacial theory applied to U-valleys,[173] advocates of the latter strongly oppose this explanation of the "U-in-U" form.[174] For instance, the statements or implications that the Alpine glaciers of the last glaciation were unable to reach the Alpine foreland, that their cross-section (according to Hess's diagrams) were only one-fifth of those of the first, or that the surface of the ice sank at each stage by about 500 m, all contradict well-known geological observations. Even the highest benches often have fresh striated surfaces, moraines and perched blocks. But it is mainly objected that the reconstructions are uncertain and illusory; that the various troughs are not discoverable except by doing violence to the evidence or by ignoring rock-structures; that some terraces are due to lateral moraines, e.g. in the Tasman Valley of New Zealand,[175] to sapping at the "black-white boundary"[176] or by bench-ice above the main glacier[177] (as in the case of the upper Aletsch Glacier to-day), to structures and rock-junctions,[178] e.g. in the basalts of the Faeroe Islands and west Greenland, in the limestone Alps of the Bernese Oberland, including those about Lauterbrunnen, to terrassing by lateral melt-water streams[179] (see p. 330), or to undercutting by a lateral glacier pressing against the flank of a trunk glacier.[180] To unite these ill-defined, irregular and discontinuous fragments or "vicinal surfaces" into continuous terraces is unjustified,[181] especially when this is done under the influence of a prevailing theory. Salomon[182] further objected that the ice, as during the Würm glaciation in the Adamello group, filled the valleys to above the shoulders, and that the terraces give more glacial epochs than can be accepted on other grounds and, being made by differential weathering, are unacceptable as old trough floors, especially in the western Alps whose inclined beds, resulting from overfolding, have enhanced the difficulties. It is denied that the earlier benches have persisted in such recognisable form or that the Alps have four troughs[183] (the Adamello group, Niedere Tauern, Norway and Altai have yielded only one[184]). It is also affirmed that the Alpine valleys had attained a low level during Mindel-Riss time (see below) and that the relative homogeneity of structure explains the rarity of these features in Scandinavia.

Lucerna,[185] from observations on the Mont Blanc region, Hohe Tatra, Liptau Alps, Karawanken and Corsica, concluded that the troughs belonged not to successive glaciations but to postglacial stages. This *reductio ad absurdum* of Hess's hypothesis, which denies any erosion to the main glaciations themselves, requires no particular refutation though others had anticipated his conclusion[186]; the evidence against it is overwhelming.

Essentially preglacial. We must reject the hypothesis that ice entirely erodes U-valleys for reasons already mentioned as well as for others now to be considered. Chief among these is the fact that U-valleys coincide with depressions excavated by preglacial streams whose courses were guided by the regional trend and inclination of the land and by the direction of

joint-systems, trough-folds or important strike or tectonic lines.[187] Bonney[188] imagined they were cut by several parallel streams and resembled straightened-out cirques (see p. 295).

That glacial overdeepening, apart from rock-basins, was merely an episode in the total history of denudation and may have been small compared with the lowering rivers accomplished is most convincingly demonstrated. Many valleys have roughly their preglacial depths; the Swiss Plain and Austrian foreland are essentially of Pliocene and pre-Deckenschotter age; Pliocene beds are found at the foot of the Italian Alps (see p. 278); preglacial lava-flows reach the present valley floors in the Caucasus[189]; and some Alpine, Pyrenean and Altai valleys were as deep or even deeper during the "great interglacial" or "great erosion" epoch[190]—the *Rinnenschotter* fills the furrows in the Swiss valleys. It is borne out by the interglacial breccias on their sides,[191] e.g. the Hötting Breccia (see p. 935); by river and lake terraces,[192] as in the Isar and Inn; by ice-moulded clefts made by streams in the lateral walls[193] or bottoms of troughs[194]; and by relics of preglacially weathered rock, occasionally discovered in these positions,[195] as in Skye and Colorado. Hanging end-moraines occur in deep lateral gorges in the Inn valley.[196] Modifications made by earlier glaciations account for the small erosion during the last glaciation.[197]

Equally strong proof is furnished by the presence of unglaciated V- and glaciated U-valleys side by side or in adjacent areas, as in the eastern Alps,[198] and by the repeated occurrence of an unaltered V-shaped river valley, which escaped glaciation, downstream from an U-shaped portion and the glacial limit fixed by outermost moraines. This has been observed in almost all glaciated lands, including Europe[199] (Alps, Black Forest, Pyrenees, Apennines, Iberian Peninsula, Auvergne), Asia[200] (Caucasus, Himalayas), America,[201] Africa[202] and New Zealand.[203] U-valleys in jointed granites, limestones, sandstones and conglomerates are found in Spain, Sicily and Greece only where Pleistocene glaciers existed.[204]

This profound contrast in transverse profiles is associated with other differences. Above, the spurs are truncated and the tributaries discordant; below, the spurs are intact and overlap and the tributaries enter at grade. Together, they show that the valleys, as in the case of the Drau, Mur and Enns, or those of the Sierra Nevadas, were virtually as deep as now before glaciation set in.[205] A like change, occasionally noticed,[206] where a glacier ascended a V-shaped valley is additional vindication.

Spur truncation presupposes pre-existing valleys since without spurs it is of course inconceivable.[207] Some valleys lay athwart the ice-flow[208] or if parallel opened icewards[209] (see below). The drift beyond the mouths is insignificant in comparison with the troughs.[210]

The view that the bigger Alpine valleys are now roughly as deep as before their glaciation is as old as the writings of Charpentier and other early glacialists. Ramsay[211] believed the ice had only moulded and slightly deepened them, a belief expressed more recently for the Alps[212] and for most parts of the world[213] (Wales, central Asia, Arran, Tauern, west Carpathians, east Pyrenees, Apennines, Vosges, Lapland, Kola Peninsula, Labrador, Sierra Nevadas of California, Andes and Kerguelen).

There was little or no modification if the ice, no matter what its depth, flowed along broad, low, mature valleys, as on the Tibetan plateau,[214] if it could not escape freely, as in the Howgill Fells and Blair Athol area of Great

Britain (see p. 217) or the longitudinal valleys of the Alps,[215] or if the valley gradient is steep,[216] e.g. in the Rhône valley below Gletsch, in Stadardalur in Iceland and in places in north Wales, the Lake District and Scottish Highlands—the valley then remains V-shaped. This was also true of valleys, main or tributary, which were transverse to the ice-flow.[217] Four classes may be distinguished[218]: (*a*) shallow valleys, with gentle slopes, glaciated by ice moving into and out of them and insufficiently rigid to span them—they are unilaterally widened, the impact side being abraded and having its angle lowered to a more or less uniform slope, the lee face suffering from plucking and depending for its configuration upon the nature of the rock; (*b*) steep and narrow, typically fluviatile valleys probably occupied by stagnant ice which was overridden along a thrust plane according to the width of the valley and the depth and velocity of the overriding ice (cf. p. 273)—it fixed the lower erosional limit and made the valleys liable to be receptacles of drift (with a certain depth the thrust plane just touched the floor) which as in Peebleshire and Selkirkshire was deposited in the lee of slopes.[219]; (*c*) valleys occupied by ice moving athwart the flow on the plateau above (see p. 273) and modified if the ice bore with sufficient pressure; and (*d*) valleys in which movement was neither transverse nor parallel but oblique.

In the first two cases, the valleys are little altered; striae are extremely rare and, like the bigger boulders of the ground-moraine, are disposed at random[220] (the latter is sometimes bedded obliquely down the valley sides[221]), while deeply decayed rock and preglacial deposits lie undisturbed,[222] as in the Adirondacks, Yukon region and the preglacial gold-bearing gravels of British Columbia. Tributary gorges, as in the Yosemite Valley,[223] remain unmodified and spurs and sinuous forms persist,[224] as in lower Glencroe (Scotland), the tributaries of the Aare, and some valleys in Labrador. Valleys with varying trends like the Tweed[225] naturally assume different forms in various parts. The contrast between the fjords and U-valleys of Norway and the valleys of east Norway that still retain much of their fluviatile character has been attributed to different relationships to ice-flow.[226]

The ice was disinclined, as early noted,[227] to move along valleys which faced it; it was, indeed, apt to stagnate in them,[228] as in the Glens of Antrim, in parts of the eastern Alps and in the Finger Lakes region (see p. 283), so that they kept their fluviatile characters. Erosion was also at a minimum where, as in the Swiss Alps,[229] a tributary impeded a major glacier or the main glacier dammed up its tributary glacier, or where, as in the Usk and certain other valleys of South Wales,[230] the valleys lay between the main glacier routes.

Glacial widening. It is implicit in the foregoing that basal sapping and lateral widening modelled the preglacial valleys which sloped and trended parallel with the flow.[231] They straightened the valleys by grinding away the ends of the spurs and cutting triangular facets. They formed betrunked hanging valleys by undercutting the base of the V in the main valley and displacing the intersection of the trunk and tributary, especially if this was steep. Orometric studies in the Tatra show that the degree of modification into troughs corresponds with the glaciation's intensity,[232] and observations in the relatively simple valleys of the Vosges give the widening as three times the deepening.[233]

The lateral erosion (Richter's *Unterschneidung*,[234] Penck's *Untergrabung*[235]) accorded with the valley's shape and the arrangement of the glacier confluents,

It made the sides concave by pressing upon their lower parts[236]—a comparison of the flaring U with the V form affords a measure of its action. Only when the V was largely transformed into the U did the lateral pressure lessen appreciably and the ice bear directly upon the floor[237] and grind out rock-basins, the erosive intensity varying with the square (Hess) or the cube (de Martonne) of the depth of the ice.[238] The U-forming erosion was stronger in flat valleys where the flow was slowed down than in steeper valleys.[239]

But overwidening without some overdeepening cannot explain all the features, especially when the tributaries were not steep.[240] Many glacialists[241] insist on the importance of vertical as distinct from horizontal erosion. Continuous trough walls which run below the level of the high-perched hanging tributaries and the gable-ends of the truncated spurs certainly vouch for overdeepening as well as widening. Advocates of overwidening are compelled to admit the erosion of the rock-basins and considerable deepening. This has been estimated at 400–500 m[242] or at *c.* 200 m or one-seventh of the total depth in some Bavarian valleys.[243] A figure of 2000 ft (*c.* 600 m) has been given for certain valleys in Alaska and British Columbia.[244] To the depth of the rock-basins, which are positively ice-eroded (see ch. XII), has to be added the lowering of the barriers below the normal "profile of equilibrium" which is much less readily assessed, though it is clear, as Wallace has shown, that the present lake *surface* and not the lake *bottom* represents the approximate level of the preglacial valley. Riegels or barriers (see p. 268) on the lateral erosion theory are due to preglacial constrictions, on the theory of vertical erosion to greater resistance.

Preglacial relief. The logical point of departure in considering this question is the preglacial relief. What, for example, was the initial surface into which the Alpine valleys of the Pleistocene were incised? If this question could be answered with any degree of assurance, the solution would be at hand. Unfortunately, the reconstruction encounters manifold difficulties. Allowance has to be made not only for subsequent ice-erosion, which has largely destroyed the preglacial surface in the French, Italian and inner Alps, but for Quaternary deformation.

Although it is agreed that the Alpine valleys already existed at the beginning of the Pleistocene, opinion is sharply divided about the stage of dissection the ice inherited. Some, following Penck, believe that the Alps' inner topography was mature[245] and had rounded watersheds and graded profiles which gave the Swiss valleys a gradient of 3–4%. This belief is based upon the sharp ascent of the Deckenschotter in the north above the valley troughs and the height of the tributaries which lie on the plane's inward continuation from the foreland. The sole of the Older Deckenschotter is seen in the region of the Rhine, Linth and Reuss glaciers[246] but in the west passes into the air.[247] The alpine region during the Oligocene and Lower Miocene had a low relief; the lower part of the Upper Bavarian Molasse, east of the Lech, and the sandstones and clays of Oligocene and Lower Miocene *Schlier* of the Alpine foreland and east Steiermark are fine grained. The marine sediments at the foot of the southern Alps are also fine-textured. The eastern Alps may have had a subdued relief[248]: it was the *Raxlandschaft*.

Nevertheless, the inner Alps, especially in Switzerland, had either a mountainous character or diverse topography[249] (Penck[250] himself recognised this for parts of the Alps); their upheaval had been too recent to have been otherwise.

Penck's view ignores the possibility of a post Sarmatic (Upper Miocene) peneplain in the Swiss and Alpine foreland and high relief in the Alps themselves. It fails to take into sufficient account the strong Alpine uplift in the late-Tertiary to which the enormous thickness of upper Pliocene bordering the range on the north, south and east testifies. This uplift produced frost at high levels[251] and steepened the floors and flanks of the valleys by rejuvenation. This is shown by the coarseness of the Nagelfluh in the Swiss Molasse, of the Upper Freshwater Molasse in Upper Bavaria, Venn basin and Tauern, and of the upper Pliocene schotter[252] and (extending the range) by the sheets of coarse detritus at the foot of the south Carpathians[253]—the highest fluvioglacial schotter in the Alpine foreland are found in valleys and not on plateaux.[254] The Bighorn Mountains, Cascade Range and Sierra Nevadas of North America, and the Himalayas, all of which display similar glacial features, were also raised in late-Pliocene time[255] (see ch. XXIX). In the Alps the upheaval continued into the Pleistocene period[256] (see p. 1326), and the changing compositions of its various schotter show that lower and lower rocks were undergoing erosion with the passage of time.[257] The preglacial downcutting has been computed at 700–800 m[258] and the late-Pliocene and Pleistocene uplifts in the central Alpine zone at 3600 m and more,[259] the amount lessening towards the margin.

Tertiary levels are extremely high in the Alps.[260] Some are extensive and flat, others are small, terrace-like ridges or remnants on the mountain crests. Rütimeyer[261] early recognised a succession of such levels. The accordance of the summits has long been known, particularly in the east where it is 1200–1400 m above the early Pleistocene floors.[262] It has been noted too in other mountains, including the Caucasus, Pyrenees, Californian Sierra Nevadas, Canadian Selkirks and Coast Range, Alaska, Greenland, Tienshan and central Asia.

This *Gipfelflur*,[263] which is most striking when the position of the erosion basis and the lithological and structural build of the mountains are most regular and constant, is ascribed to planation by frost[264] or interpreted as the upper denudation level above which mountains cannot grow.[265] Some believe it results from forces now at work[266] while others, who think there is no upper limit of denudation, deny this.[267] Others again deem it to be an ancient surface[268] or combine this with an upper denudation level.[269]

The Alpine *Gipfelflur*, which according to some is tectonic[270] and closely related to the lithology, is in Heim's opinion[271] unconnected with geological structure, except locally. It has been inherited from *Hochfalt* forms by an interaction of denudation and isostasy, being related to erosion-basis, valley density and the angle of valley sides. Mountains of a structural height of 50,000 m have, he thinks, been reduced to a plain during and since the uplift. Other morphologists agree that it is independent of rock-structure but regard it as a planation surface.[272]

The *Gipfelflur* is now known to be composite,[273] the sum of several morphological surfaces (*Niveaux*; *Flächensysteme*; *Stockwerke*) of various ages. These elements, which have been influenced only subordinately by selective erosion,[274] have been lowered into each other as a *Gipfelflurtreppe*; the highest *Flur* of all, Richter's real (*echte*) *Gipfelflur*, is itself possibly only a remnant cut out of a yet older surface: in Tyrol[275] the vertical distance between the *Gipfelflur* and the middle Tertiary (Miocene) surface, the oldest surface still widely preserved throughout the Alps, varies between 100 m (Kitzbühl

Alps) and 1100 m (Oetztal Alps). Each surface with its outliers (*Restberge*) projects from the lower plane and falls outwards from the central Alps to north and south, being traceable throughout the Alps from east to west.

Creutzberg's *Firnfeldniveau*[276] (Lautensach's *Bedrettotalboden*,[277] *Petanetto* Terrace of Tessin region and *Simmenfluhniveau* of the Bernese Oberland[278]) which lies below the summit level and the middle Tertiary surface and the steep-sided ridges,[279] is a wide landscape to which unequal uplifts have imparted quite substantial differences of level[280]; in places in the Alps, it appears to be two surfaces.[281] In the central Alps, it often carries firnfields, whence its name (other levels also have firnfields[282]), as was recognised by Richter[283] who erroneously ascribed it to nivation (see p. 307). Occurring at about 2000 m, it is well seen in the east, north and central Alps[284] and may have a representative in the Swiss Alps.[285] Its date is generally accepted as early Miocene[286] (it has been placed in the Pontic[287] or extended backwards into the Oligocene[288]) since beds of this age rest upon it near Gratz and fresh-water beds with congiomerates, plants and coal seams of lower Miocene age occur at the eastern part of the south Alps and in the southern Limestone Alps, as in the Klagenfurt basin and Levant, Mürz and Enns valleys. There is no longer any trace of the lower Miocene terrace in the eastern Alps[289] (see below) where the oldest surface is middle Miocene. In the higher and more dissected western Alps it is above the highest summits.[290]

After making the *Firnfeldniveau*, the destructive processes never again established flat surfaces of any appreciable width,[291] though an occasional lower level has been distinguished[292] such as Creutzberg's *Hochtalboden* (Lautensach's *Sobriotalboden*[293]) whose age is variously regarded as upper Miocene[294] (Pontic), upper Pliocene[295] or interglacial[296] but may be lower Pliocene.[297] Remnants of earlier and higher surfaces are preserved as valley floors, low graded valley sides or mountain ridges (*Eckfluren*[298]): they add to the confusion of levels. A final preglacial surface beneath the Deckenschotter is also often recognised.[299]

Beck[300] recognised four levels in the Bernese Oberland and styled them *Simmenfluhniveau* (1500 m), *Burgflurniveau* (1000 m), *Kirchetniveau* (700 m) and *Hilterfingenniveau* (600 m); the first two preceeded the Deckenschotter glaciations and marine Pliocene on the southern margin of the Alps.

In the eastern Alps, where much recent work has reconstructed the preglacial relief and where the waves in the *Gipfelflur* are less pronounced and the remnants of the older surfaces consequently better preserved, the successive levels have been deemed to be peneplains[301] formed in times of tectonic rest and uplifted at dates not yet definitely ascertained. Schwinner[302] detected four such erosion cycles (lower Miocene, Pontic, lower Quaternary and Würm). Machatschek[303] recognised a highest surface of early Tertiary age and later ones of lower and middle Miocene, upper Miocene, late-Tertiary and preglacial age. Rinaldini[304] held similar views. The majority put the oldest surface in the lower Miocene, though some have thought no surface older than upper Miocene or lower Pliocene is now preserved.[305] Others prefer to regard the different features not as relics of valley floors but as tangents of the highest levels of the old surfaces.[306]

Winkler,[307] followed in general by E. Seefeldner,[308] thought the determining factor was the shifting of the basis of erosion by eustatic transgressions (of lower Miocene and lower Pliocene age) which were periods of

planation with periods of regression and rejuvenation (upper Oligocene, base of Miocene and base of Pliocene).

The levels of the various regions differ considerably in height as seen in R. Brinkmann's map of the *Gipfelflur*[309] or v. Klebelsberg's figures for Tyrol.[310] The difference has been attributed to selective erosion[311]; to original differences in the mountain building levels; to folding[312] or *Grossfaltung*[313] or uplift as a whole[314]; to varying distances from the erosion bases of the Alpine margin or large valleys[315]; or to the shifting bases of erosion or eustatic trangressions and regressions.[316] There is however no agreement upon the number of levels.

Thus the existing relief appears to have been created by young crustal movements. They succeeded the principal folding and had a preponderating vertical component, due either to block faulting, as in the central Alps (where faults with throws of several thousands of metres occur), or to geanticlinal warping (*Grossfaltung*), accompanied by transverse and longitudinal flexures.[317] In the central Alps, the *Gipfelflur* is 800 m higher than on the southern margin of the Limestone Alps.[318] It rises in the Zillertal to 3500 m and in the Oetztal Alps to 3700 m but sinks to below 2500 m in the Inn valley.[319] The association of profound glacial modification with such uplifted regions and rapid ice-flow and the milder effects noticeable where the *Gipfelflur* sagged, explain quite naturally why glacialists who have studied one Alpine region are glacial erosionists and their colleagues in others defend a glacier's incapacity to erode.[320]

A young Pliocene doming of the south Alps was shown by Penck, Brückner and Stefani, and for the eastern Alps by Winkler[321] while numerous writers have stressed the importance of late tectonic movements.[322] This *Grossfaltung* (undation), which was steeper towards the margin and may be isostatic following denudation,[323] is supported by N. Krebs, F. Leyden and F. Machatschek. It is confirmed by the isogams and distribution of gravity[324] and by the distribution of ore-deposits.[325] It was thought by W. Penck to have been continuous but by others discontinuous.[326] The *Gipfelflur* may have been lifted 1800–2000 m.[327]

As noticed already, opinions differ as to the number of movements that have elevated the Alps; three,[328] four[329] or five[330] rejuvenated phases have been recognised. The discrepancies, due partly to the entry of subordinate stages, will only be eliminated by careful mapping on lines already initiated.[331] Great care has to be exercised in combining the isolated remnants into systems[332] and in unravelling the complications introduced by glaciation and by tectonic causes, including continued undation, that have deformed the surfaces.[333] Uniformity is perhaps, in any case, not to be expected in a structural system so complicated as the Alps.

In fine, the present drainage lines of the Alps are inherited from Tertiary ones at higher levels. They coincide with downwarps of the *Gipfelflur* which has swells and sags superimposed transversely and longitudinally upon the main doming.[334] The rejuvenated rivers have eaten back into the mountains in a series of curves dissecting earlier features because of successive uplifts which resulted from a steep, marginal flexuring (see p. 1326). Confirmation is provided by the youthful stage of the earth-movements, as in the Alps,[335] and by the absence of U-valleys and hanging tributaries in regions like Tibet[336] or parts of the Caucasus[337] which were untouched by preglacial uplift and dissection in accord with D. W. Johnson's general statement.[338]

Such a sequence of uplifts and erosion cycles has been discovered in the Isonzo region[339] and in the Carpathians[340] which, compared with the Alps, have better preserved the impress of the phases of maturity and have been less touched by ice-erosion.[341] The relief was immature not only in the Alps[342] but in the Pyrenees (Oligocene, Miocene and Pliocene)[343] and Hohe Tauern,[344] Hohe Tatra,[345] Apennines,[346] Pyrenees,[347] British Isles,[348] Norway[349] and North America.[350] The Alpine rejuvenation is certain: the Alps are too high to have had graded rivers throughout; the valleys grade on the north with their relatively low peneplain of central Switzerland, on the south with the *Senkungsfeld* of Lombardy; the preglacial schotter occupies a low position at the embouchures on the south and east; the U-valleys, as in the Limestone Alps,[351] pass downstream into rejuvenated valleys (see p. 322) which notch the shoulders of the high-level surfaces, while typical hanging valleys occur in glaciated regions like the Maritime Alps and glaciated and unglaciated valleys differ little in depth,[352] as in the Limestone Alps and eastern Alps generally. Had the Alpine valleys not been rejuvenated, it would be necessary to assume with Hess that their floors lay very high and fell abruptly at the Alpine flanks.

Steps and rejuvenations. Steps mark the confluence of valleys, either in the trunk *Talweg* or at the *Talschluss*. Crowned often by rock-barriers, they form a vital link in proving rejuvenation. Thoroughgoing glacial erosionists[353] associated them with a sudden increase in the vertical erosion at the point of confluence, either of glaciers or of firn masses, whence the term confluence steps. This is proved by soundings in the Concordiaplatz of the Great Aletsch Glacier (see p. 37) and harmonises with the law that the sum of the cross-sections of two streams is more than the cross-section of the confluent stream.[354] Nevertheless, steps are absent from many confluences[355] (e.g. Oetztal and Inntal, Zillertal and Inntal, Mölltal and Drautal), and conversely occur where there was no confluence[356] or outside the direct path of the ice.[357] Others are "selective" or "resistance" steps,[358] made by hard rocks or resistant structures[359] (the association, however, is not invariable[360]), by cross faults,[361] or by variations in the pattern or strength of the jointing.[362] Such steps, being rooted in position, are presumably somewhat stable features in the glacial cycle.

While some believe the steps were created by ice[363] and were not inherited from an older relief, most geologists[364] share the opinion of de Martonne and Sölch that the transverse and longitudinal profiles were discontinuous and that the steps, no doubt accentuated and sometimes originated by confluence, were not usually created in this way. Whether they are *Talquerstufen*, *Mündungsstufen* or *Trogschlüsse*, they were initiated by preglacial,[365] interglacial[366] or glacial[367] uplift and rejuvenation.

That the heads of fluviatile erosion were the forerunners of the glacial steps is suggested by the following: the steps are of great height, e.g. 700 m in the Tauern,[368] and correspond in trunk and tributary valleys and with the benches[369]; their distance up the tributaries decreases the higher the confluence of the main and tributary valleys[370]; and they are equally well developed in valleys that were not glaciated[371] (Alps, Black Forest) or suffered little glaciation[372] (Andes). Penck[373] himself recognised the glacial theory's inability to account for the sudden increase in erosion at the lateral steps and *Talschluss*. Mature valleys, as in the Taurus, north Caucasus, west Turkestan and Karakoram Mountains, have neither steps nor hanging valleys.[374]

Ice exaggerates any inequality in its bed[375] by causing frost action[376] due to varying pressure (see p. 301); by plucking from the face of the rise[377] (this squares off the upper part and differentiates it from the typical cirque back); by special weighting as it pushes forward over the brink and before it breaks off and drops down[378]; and by basal sapping in transverse crevasses where these, at the coming and going of glaciers, extended to a rock which had no protective drift and was subject to oscillations about freezing point,[379] conditions not unlike those at the bottom of the bergschrund (see p. 299)—Hess[380] associated the *Trogschluss* with the bergschrund. Nivation at the risers acted at the advent of each glaciation.[381]

While the ice flattens the treads or even hollows out rock-basins (the contention[382] that it cannot excavate these on a uniform incline conflicts with the view[383] that they arise from variations in a valley's width and the entrance of tributary glaciers) it heightens the rise by its recession; in this way, steps resemble huge roches moutonnées,[384] and every gradation may be found between small roches moutonnées to steps which occupy the whole width of a valley floor.[385] De Martonne (see p. 269) affirmed that erosion increased with the gradient up to a certain limit, placed by Russell[386] at about 30°. Beyond this it abated since the pressure is proportional to the cosine of the declivity and the ice comes less intimately in contact with its bed. This may be true for abrasion but plucking is favoured in just those positions of excessive gradient (see p. 251). Plucking also vitiates the deduction[387] that erosion diminishes as the slope increases.

That preglacial rejuvenation reacted in this way is denied, mainly by American geologists.[388] Uplift, they assert, induces only minor gorges; discordance is ephemeral, is present only during early youth and is incompatible with the breadth of a typical U-valley and its steps; the youthful age may be elided if uplift is slow, and there may be no marked discordance even if the uplift is rapid, unless pronounced tilting accompanies it and greatly accelerates the trunk stream, leaving branches unaffected; the tributaries were entrenched as gorges in the lateral scarps; the rejuvenated valleys were probably so narrow that the conversion of a V into a U involved much excavation, and if wide and deep the walls would have been broken down simultaneously; and a revival of normal erosion in all mountains shortly before their glaciation is very unlikely.

These objections notwithstanding, it seems that U-valleys were river valleys preglacially rejuvenated. Their features prove this inheritance: the bench represents the topographic shoulder, the steps the limits of the various rejuvenations and the *Trogplatte* the upper part of an older floor. Glacial widening and overdeepening modified the valleys more or less severely—in the mature or senile stage of broad valleys there would be only moderate opportunities for marked local increase of ice-velocity so that the profiles, even after glaciation, would still be largely at grade[389] (see p. 328).

There remains, as in the case of the cirques and rock-basins, some uncertainty as to the precise way in which this was done. The ice remodelled and fashioned the U-form by rising at its margin[390] (see p. 103) and by undercutting,[391] as Richter maintained and as observations on modern glaciers suggest (see p. 318), at the *abrupts de sapement*,[392] analogous with the undercutting by streams on the outer sides of curves. The dome shape caused the ice to press down on the sides rather than on the floors[393] (see p. 324). Frost sapped the walls along the black-white frontier[394] in conditions similar to

those at the cirque wall (except in the direction of motion), its action being helped by block movement (see p. 118), by marginal crevasses, by abrassive material supplied from the overhanging walls[395] (which on account of the thinness and rigidity of the ice was firmly frozen into the ice), by seepage from the sides[396] and by marginal melt-waters[397] like those now active on the glaciers of Zemmgrund (Oetztal) and Argentière.[398] Sapping, however, was probably less than plucking, especially if the walls were jointed, the ice was thin, and fragments were able to penetrate through crevasses to the base[399]; for crevasses seldom reach this (see p. 36) as Johnson[400] himself recognised. Preglacial or interglacial landslipping may have facilitated widening. It has been stated that glaciers which progress by block-movement (see p. 118) erode more intensely than those which have a continuous flow (see p. 215), and that consequently the troughs are mostly due to ice-erosion during the early stages of glaciation.[401] Above the ice itself, nivation and sand- and snow-blast were active[402] (see p. 302).

Autogenetic step-development by accentuating, through sapping, the initial breaks of slope causes the steps to retreat upstream as long as sapping is active. As a corollary of this theory, G. Taylor,[403] forced by the necessity of dissecting an upland by glacial action alone working on an initial landscape of tectonic origin without a previous history of normal erosion, advanced the view (already suggested for some hanging valleys[404]) that cirques by headward glacial erosion excavated U-valleys—some hanging valleys in the Alps are significantly present only on slopes facing north or east.[405] This action however, was quite inadequate, unless possibly the floor was tilted[406] or plucking brought about the transformation.[407] The Antarctic valleys in question were more probably formed tectonically.[408]

Some writers,[409] emphasising that a glacier bed carries water as well as ice, stress the power of subglacial streams operating conjointly with the ice and being repeatedly deviated by it. Although V-shaped trenches in old glacier-beds, like those which notch rock-barriers or that retreating modern glaciers have exposed,[410] are cited as proof, they are too infrequent to make this action of any moment.[411]

Erosion by lateral streams. Lateral streams, working under a glacier's peculiar control, either submarginally or between the ice and its wall, have been held responsible for spur truncation and widening.[412] They incise two lateral and parallel canons (*double canon glaciare* of Brunhes) in which pot hole drilling is important,[413] while plucking and abrasion lower and remove the central rib and widen and round the river furrows. Brunhes[414] repeatedly advocated this simultaneous erosion by ice and lateral streams. His view has won considerable approval[415] and been extended to cirques.[416] Streams working on the sunny side may have produced the asymmetry in longitudinal valleys in the Himalayas.[417]

In vindication of Brunhes' view have been cited the sporadic depressions (*gouttières*[418]) along the valley sides,[419] and river channels flanking U-valleys[420] as in the Etsch and Rhine and about the Upper Grindelwald Glacier—talus may obscure others. Suggestive too are the rounded masses (*Platten*; see p. 269) which project upwards in front of modern glaciers and divide their waters into two or more streams,[421] e.g. in Alaska and at the snout of the Upper Grindelwald, Fiesch and Übeltal glaciers.

This action has undoubtedly played a minor role.[422] But, as a major factor, it must be rejected since lateral streams, associated with modern

Alpine glaciers, are exceptional and not limited to two[423]; signs of fluviatile erosion in abandoned glaciated valleys are rare, especially on barriers and steps; crevasses leading to subglacial water-channels are restricted in general to glaciers not thicker than 60 m[424]; and some trough-shaped valleys were above the glacial snowline where there were no streams.[425]

Interglacial streams and ice-protection. Many glacialists, starting from the double view that rivers erode and glaciers protect, associate the over-deepening with torrents discharged in the main valley from receding snouts during warm interglacial epochs when lingering glaciers protected the tributaries at higher and colder levels. Glaciers covered and shielded the landings above the steps in trunk valleys while melt-waters eroded the treads below. Thus the several steps mark the upper limits of regressive erosion by interglacial streams and testify to the slightness of direct glacial erosion.

This view, presented by Garwood[426] from observations in the Alps and Himalayas and arrived at independently by W. Kilian[427]—he later abandoned it in favour of rejuvenation[428] (see below)—has been advocated for various parts of the world.[429] Heim,[430] on the evidence of bores in the Rhine and Aare,[431] relegated the downcutting in Swiss valleys, amounting to as much as 1000–2000 m,[432] to the great interglacial epoch. Continuous uplift accelerated the river's tempo.[433] According to Machatschek,[434] the Rhône valley near Visp was lowered by 600 m during the first two interglacial epochs (principally during the second) and by only half this amount during the first two glaciations. The lowering during the third interglacial was 150–200 m. Of the total lowering, therefore, 750–800 m was accomplished by interglacial erosion (made possible by tectonic uplift of the Alps) and only 400–450 m by glacial erosion (see p. 321).

Since the interglacial epochs were much longer than the glacial epochs (see p. 918), it goes without saying that the interglacial and to a far less degree the interstadial streams had their share in the lowering: the stepped *Talschluss* has been attributed to ice and water working alternately.[435] U-valleys, therefore, were cut partly during the advance and retreat of the glaciers and were exposed alternately to epiglacial cycles of river and ice erosion[436]—they may in this way have been lowered 430 m, of which 80 m was interglacial,[437] or 1000–1200 m.[438] Difficulties enter when we attempt to apportion the parts played by the ice of the glacial epochs and the rivers of interglacial and preglacial times. While the glacial erosionist probably underestimates the latter, advocates of Garwood's view and of the multiple trough hypothesis (see p. 320) take insufficient note of erosion by preglacial streams[439] since preglacial time much exceeded the longest interglacial epoch.

Garwood's hypothesis has little to commend it though interglacial stream gorges, confined usually to rock-barriers, do occasionally occur.[440] It supposes, contrary to general opinion (see p. 919), that the first interglacial epoch in the Alps saw the greatest retreat[441] and that glaciers survived during the interglacial epochs. It ignores the drift which filled the interglacial valleys and impeded active valley enlargement, so that this was much less than the sum of the interglacial durations might suggest. It is inconsistent with many other observations. For example, hanging valleys are frequently U-shaped and have rock-barriers and basins[442]; there are a great number of small steps[443]; moraines are frequently wanting at the steps[444]—subsequent removal[445] is unlikely; trough-floors are wide and flat; streams from hanging valleys have not established accordance[446]; troughs cease where glaciers end

(see p. 322); and the steps in the main and tributary valleys are not systematically related in number, height or spacing.[447] It is, moreover, improbable that the glacier snouts were nicely poised just where the tributaries and main valleys were confluent[448] or, to judge from the behaviour of modern glaciers (see ch. VI), that they stood in these positions without oscillating.[449] Furthermore, aggradation rather than erosion takes place below glacier snouts[450] and interglacial rivers, like those of to-day (see ch. XXV), tended to fill rock-basins, erode barriers and eliminate breaks in the profile.[451]

The gravest objection concerns the postulate that glaciers lingered longest in the tributaries. This is contrary to observations on modern glaciers[452] including the Durance valley where Kilian worked out his theory. Main glaciers rooted in large névés respond to climatic changes less rapidly than do smaller ones. It conflicts too with the countless instances of marginal drainage and the presence of hanging valleys where the main valley only was glaciated and the lateral valleys had no glacier protection.[453]

Finally, interglacial epochs in the heart of such regions as Scandinavia and Scotland which display typical U-valleys, though most probable, have yet to be conclusively demonstrated.

Transfluence and diffluence. Branches of glaciers, if thick enough, overtop divides and sweeping over these terminate in pendant masses or coalesce with other glaciers. They lower, widen and round cols (Kilian's *seuils de débordement*[454]), striate the sides horizontally, scoup out pass-lakes (sometimes with outlets at either end) and transform the cross-section into a U. Instances of such lowered cols or U-passes are some of the Scottish passes,[455] e.g. the Pass of Drumochter and the Loch Treig col in the Scottish Highlands which may (doubtfully) have been lowered 450–550 m[456]; Alpine passes,[457] e.g. Mont Cenis, Simplon (2009 m), Grimsel, Gotthard (2112 m), Bernina (2330 m), Splügen (2117 m), Maloja (1817 m) and Brenner (1370 m; this U-trough is about 300 m deep), those across the North Limestone Alps[458] and in the Eastern Alps; the *skar* through the main Scandinavian divide[459] which the eccentric ice lowered and those across the heads of the Finger Lakes.[460]

Diffluence occurred if the ice near the glacial periphery overflowed in sufficient strength and duration.[461] Diffluence cols[462] were sometimes lowered and widened—in the initial stage possibly by overflow waters from ponded glacier-lakes[463]—to become the main lines of ice-discharge.[464] They were indeed sometimes wholly removed and replaced by U-shaped through valleys,[465] with high, steep and truncated walls, low flat divides, broad and rounded, and studded with pass-lakes or swamps. Examples occur in all glaciated countries, for example in Labrador, where the U-valleys are athwart the high, interior range,[466] south of the Finger Lakes, between the St. Lawrence and Susquehanna rivers,[467] and in Scotland—the Bhealaich of the Highlands. Some through valleys may be preglacial like the Tarbets of the Scottish Highlands while others may have been hollowed out by valley glaciers flowing in opposite directions from the same source—the modern glacier-filled through valleys are the through glaciers (see p. 88).

Remnants of the watersheds may rise as *Inselberge* or islands out of the pass-lakes, such as Monte Isola in Lago d'Iseo or the Burgenstock in Lake Lucerne,[468] though *Inselberge* may also be made by lateral streams (see p. 330) or by truncating lateral spurs.[469]

Diffluence has been credited not only with lowering cols, eroding "dif-

fluence basins",[470] and the forking of such Alpine *Randseen* as Como, Lugano, Iseo and Garda, but with the bifurcation of certain Alpine U-valleys,[471] e.g. the Rhine near Sargans, the Isère at Montmélian and the Salzach near Zell, the creation of "diffluence steps" west of the cols on the Norwegian watershed,[472] and of "diffluence spurs" in the angle between the overflowing glacier and main valley.[473] It has been held responsible for the depressions that margin the bosses before the Jostedalsbrae[474] and for some of the Norwegian straits.[475] In few places of diffluence was the accordance of level between the floors of diverging tributaries close enough to give a perfect bifurcation like that of Lago di Como where the two glacial outlets have been of approximately equal cross-section.

Three types of glacial pass have been recognised[476]; the first is the modified cirque pass of the iceshed or the Törl type of the eastern Alps; this merges by rounding into the second or *Karjoch* type; and this by further erosion develops into transfluence and diffluence passes where, for example, the firn of a higher cirque spills through a col into another firn at a lower level.

Glacial modification has in this way sometimes brought about glacial capture as Davis[477] deduced theoretically. This seems to have been first recognised by S. Meunier[478] who suggested that the drift succession in Switzerland resulted possibly from the enlargement of glaciers by the capture of adjoining reservoirs. Matthes[479] noted the first example, in the Bighorn Mountains, and others have been observed[480] in the Alps and elsewhere. M. Conway[481] thought that glaciers might by frost action eat through a range and draw off part of the higher ice. Glacial capture may also have been due to tectonic influences[482] or to interglacial rivers.[483]

Conclusion. The present relief is the outcome of a prolonged development, the beginnings of which go far back into the Tertiary era. The Yosemite Valley, California, monographed by Matthes,[484] is perhaps the world's finest U-valley. This gigantic chasm, with cliffs rising 3000–4000 ft (914–1220 m) above the floor, epitomises and illustrates in its literature the change of opinion concerning such valleys; it has been referred to faulting,[485] to preglacial streams working along joints[486] and to ice-erosion.[487] Yet Matthes proved that its history is that of a rejuvenated canyon gravely modified by ice and over-deepened by *c.* 450 m. This valley embodies the compromise between ice and stream erosion, now seemingly established for U-valleys in general. The ice accentuated rather than eroded or conserved. V-shaped valleys, the outcome of fluviatile cycles, were more or less seriously altered by glaciers which flowed through them. The glaciers straightened and steepened the valleys by truncating the lateral spurs, so accentuating the difference between the shoulder and the slopes. They exaggerated the irregularities of the floor, hollowed out rock-basins, heightened the rise of the steps and converted the uppermost into the *Talschluss*. The overdeepening was sometimes quantitatively profound; rock-basins represent not its full measure but the excess of the lowering in the basins over the lowering on other parts of the floor.

Penck,[488] the leading European protagonist of the glacial theory, recognised that glaciers operate mainly in modifying pre-existing features created by running water; the typical U-valleys of the Alps, for instance, are restricted to the raised *Grosssattel* region. The Colorado Canyon, if glaciated, would present all the features of a typical U-valley.[489]

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Chapter XV

FJORDS, FJÄRDS AND FÖHRDES

1. *Fjords*

Form. The words *fjord* in Norway, *fiord* in Sweden, *fjordur* in Iceland and the Faeroes, *förde* in Denmark and *föhrde* in Germany are all derived from the same root. J. D. Dana,[1] who was the first to recognise the fjord as a special coastal type, separated the first from the others as a technical term.

Fjords, which constitute one of the best-defined coastal types, everywhere exhibit the same general and essential characters.[2] They are long, straight and relatively narrow sea-inlets (the Sognefjord whose trough extends from 900 m or 1200 m above to 1200 m below sea-level, i.e. 2100–2400 m, is 230 km long and only 4·8 km broad[3]) penetrating far into a mountainous land. Their walls are high and regular, parallel or subparallel, and without conspicuous rock-spurs. Sometimes, as commonly in Greenland and Patagonia, they pass at their heads into supramarine U-valleys, with or without elongated rock-basins or moraine-dammed lakes: the term fjord has, indeed, been extended to embrace these "fjord valleys" with their "fjord lakes".[4] In other places, as in the Lofoten Islands and Lysefjord, Norway, the walls close in and the fjord ends abruptly without a fjord valley. Such "sackfjords" (*sekkefjorder*) as Helland[5] named them (they are Nansen's "cirque-fjords" or "cul-de-sac fjords"[6]) characterise north Norway but occur in other lands, e.g. Tierra del Fuego.[7]

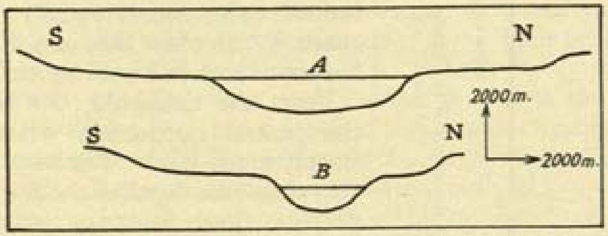

Fig. 60.—Transverse profiles of the Sognefjord. A. Between Fuglesaetfjord and Vadheimfjord. B. Over Frönningen. H. W:son Ählmann, *18*, p. 111, fig. 50.

The fjord's sides are steep both above and below the waterline. Though sometimes vertical they slope as a rule at not more than 30° or 40°[8] and give trough-like sections (fig. 60). Their heights above the sea may attain exceptionally 1525 or 1830 m as in south-east Alaska[9] or 1800 m in Franz Josef Fjord, east Greenland[10]—this highest marine cliff in the world, with the adjacent depth of more than 400 fathoms (732 m), gives the exceptional "vertical extreme" of *c.* 2530 m. Tributary valleys, ranging from small cirques to big, spurless troughs, hang and their streams plunge in a series of cascades or in one great leap into the fjord below. The floors of the lateral fjords may also be higher than those of the main fjords,[11] as in British

Columbia, Alaska, Greenland, Patagonia and the Sognefjord, the difference here being as much as 1000 m.

The most important element in the fjord problem is the threshold, the *eide* of Norway (figs. 61, 62), which Captain J. Cook[12] first discovered in Tierra del Fuego (Cook Bay) and New Zealand (Dusky Bay) and was familiar to early South American sealers.[13] The threshold may be above or at the mouth and a single fjord may have several shallow parts. In Greenland, the thresholds are rarely covered by more than about 200 m of water[14] and in Patagonia by 100 m.[15] Lateral fjords also commonly have thresholds.[16]

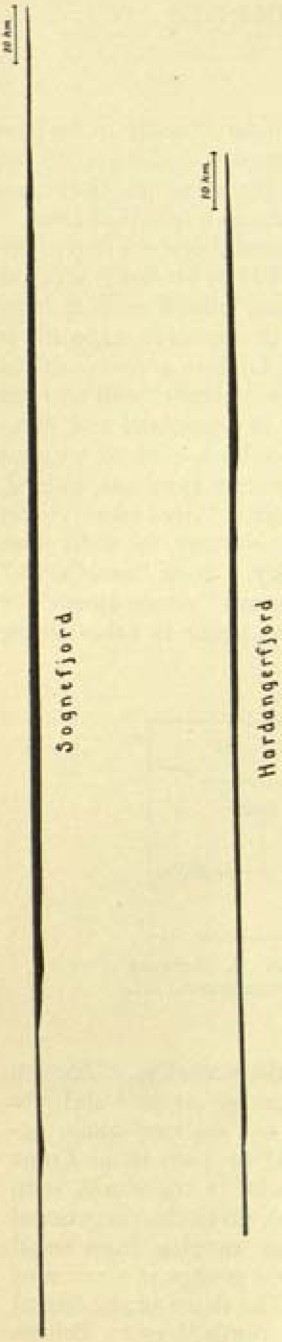

FIG. 61.—Longitudinal profiles of the Sogne and Hardanger fjords, Norway. N. H. Kolderup, *N. G. T.* 12, 1931, p. 455, fig. 6.

Thresholds, which create uniform conditions and even stagnation in the basins behind them, are essential features in fjords[17] (inlets without them have been called pseudo-fjords[18]). Yet a few geographers,[19] while considering them to be usual, deem them to be unessential; about one-fifth of the Scottish sea-lochs have no threshold.[20] Fjords differ in this respect from the wedge-shaped inlets or *rias* which in unglaciated lands like north Spain (*ria*, a river mouth, is local here), Brittany, south-west Ireland, Corsica and south-east China, gradually widen and deepen seawards.[21] The *ria*, which Richthofen[22] established as a distinctive type, is a submerged, longitudinal valley found usually on transverse coasts.[23] It often lies in softer rocks and has more rounded outlines than the fjords.

From the threshold, the floor sinks to the central fjord-basin whose sides rise steeply to sea-level. The basin is sometimes very deep; the depth of the Sogne, Norway's deepest fjord, is 1244 m and is only exceeded by Baker Fjord (1261 m) and Messier Fjord (1270 m) in Patagonia[24] and North-west Fjord (793 fathoms; 1450 m) in Scoresby Sound—Franz Josef Fjord is 546 fathoms (998 m) deep. The greatest known fjord depth in Alaska is 883 m (in the outer part of Chatham Strait) and in British Columbia 785 m (in Finlayson Channel). In Loch Morar, the deepest of the Scottish lochs, which only just escapes being a true sea-loch (its surface is 30 ft or 9 m above the sea), the floor is at −301 m (987 ft) O.D., a depth not equalled in the Atlantic Ocean for some 120 miles

(*c.* 200 km) west of this point. Dinse[25] has given the dimensions and depths of the world's fjords.

Although a fjord, as in Loch Etive and Loch Fyne in Scotland, may be co-extensive with its basin, it has customarily several "basins of the second order" as Dinse named them. Of varied size and shape, they resemble those of the supramarine fjord valleys. There is no order in their relative depths or positions; the deepest basin may be mid-way or near the mouth or head. Yet each basin is usually shovel-shaped and deepest at its inner end though exceptions to this rule are numerous, as in Scotland where as many simple basins are deepest at the outer end.[26] Sometimes, as reported from Alaska,[27] the greatest depth is to one side and due to erosion working down a dip slope.

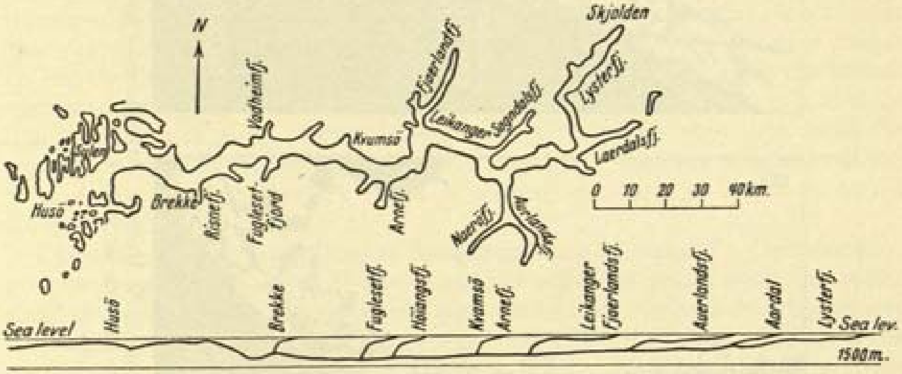

FIG. 62.—Map of the Sognefjord with longitudinal profiles of the main fjord and tributaries. H. W:son Åhlmann, *18*, figs. 71 and 134.

It is necessary, as in the case of the rock-basins in U-valleys (see p. 263), to consider the basins in natural proportions (fig. 61). Thus the Sognefjord[28] sinks from the upper end at an angle of 0° 39′ to 1244 m, 25 miles (40 km) from the entrance, and then rises with a slope of 1° 2′ to the threshold at 158 m. The Hardangerfjord has a depth of 800 m and a threshold at 175 m. In general, the bottom of the Norwegian fjords falls towards the mouth at an angle of 0·5–1·5° and then rises somewhat more rapidly at 1·5–2·5°,[29] though they sometimes shoal much more suddenly. It is manifestly the great length which makes great depths possible and explains why the longest fjords are usually the deepest[30]; the relatively short Icelandic fjords descend only to *c.* 100 m[31] and the smaller of the west Greenland fjords to 100–200 m as against the 200–400 m or even 700 m of the larger fjords[32] and the 1055 m of the Upernivik Icefjord.[33]

Fjords, as Dana and Peschel remarked,[34] occur not singly but in groups. Their members are parallel among themselves[35] and if they change their course or receive a tributary fjord they do so through a high angle, frequently a right angle[36] (fig. 63). They are directly connected by fjord straits[37] and outside their mouths, where the rocks are resistant,[38] there are belts of islands, the skerry guard (Norw. *Skjaergård*), as off the west coast of Norway, Scotland (Lewis is the analogue of the Lofoten Islands), Greenland and Patagonia. Iceland has no such feature.[39] Skerry guards are separated from the mainland by *Längsfjords*, parallel with the coasts and athwart the fjords, e.g.

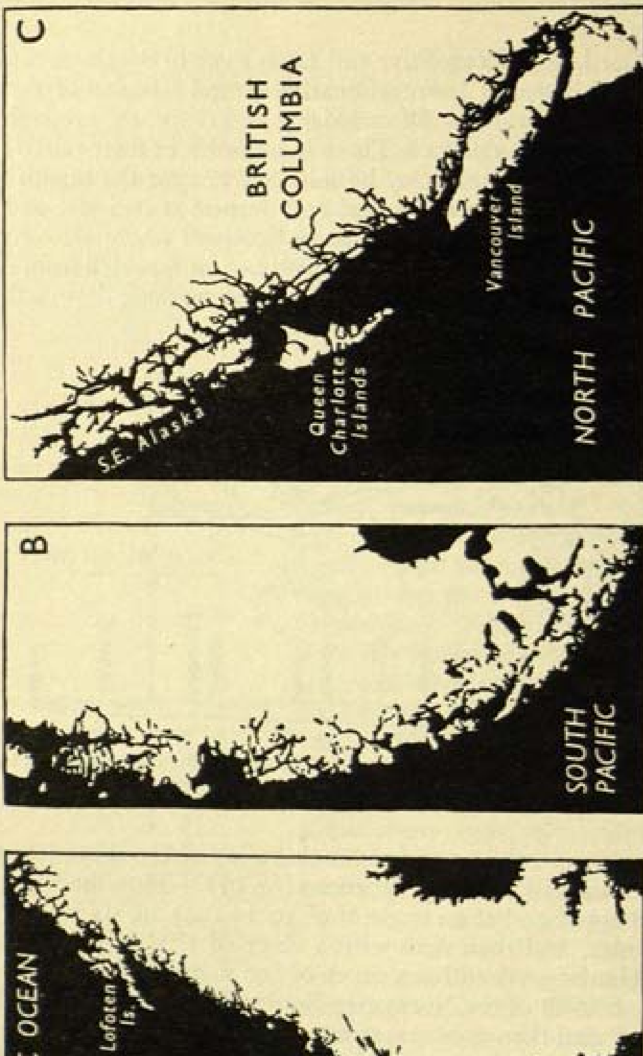

Fig. 63.—Three types of fjord coast. A. Holmes, *Principles of Physical Geology*, 1945, p. 225, fig. 117.

Norway's Trondheimled and Vestfjorden and the Minch of Scotland. Such passages are less well developed in Greenland or New Zealand.

A fjord coast is highly indented; New Zealand's is estimated at 2366·5 km (mainland, 1582·7; islands, 783·8) and Norway's, between Silde Gabet and Fensfjord, at 5421·1 km (mainland, 2916·6; islands, 3224·6). The fjord coasts, of the world were computed in 1903 to be 700,000 km long or 20 times the length of the smooth coastal curve. This figure is, however, much too high though no later determination has seemingly been made.[40]

Relation to rock-structure. Fjords predominantly occur in massive crystalline rocks, such as granites in west Greenland and north Norway, diorites and granites in New Zealand, granodiorites in Patagonia, or early Palaeozoic rocks in Norway. But typical forms are found in other kinds, in Archaean rocks in Scotland and in Mesozoic and Tertiary strata in Alaska and Spitsbergen. They are well displayed too in the Tertiary volcanic plateaux of Iceland and Kerguelen, in folded mountains and longitudinal coasts, as in Patagonia, British Columbia, Alaska, New Zealand and Norway (N. of 62° N. Lat.), in the faulted coasts of Scotland, Norway and Greenland, and in the *Rumpfgebirge* of Greenland and Norway (S. of 62° N. Lat.). They are rare and small in the rigid Pre-Cambrian foreland of Norway.[41]

The shape varies considerably with the rock. Thus fjords in Norway's gneiss and granite are narrow and steep-walled[42] while in the sedimentary terrain of Greenland, Finnmark and Spitsbergen, they are wider, less regular and not so steep-sided,[43] and contract if they pass through massive rocks. In the basalts[44] of Iceland and of west and east Greenland, they are broad, open and generally short. Islands occur where soft rocks lead to a widening.[45] Fjords in parts of east Greenland shallow where they pass from the gneissic tract to the folded zone.[46] In Patagonia,[47] they are absent from the horizontally bedded Tertiary rocks or assume the form of fjords or broad *Zungenbecken*.

Fjords are usually transverse to the general coastal trend.[48] Occasionally, as near Bergen in Norway,[49] they coincide with the grain, with the softer rock, or with synclinal axes, e.g. the east–west fjords between Nordfjord and Sognefjord though the latter cuts across faults and rock-contacts and is the outstanding example of a superimposed fjord. They also lie along the boundaries of different kinds of rock[50]; the Hardangerfjord parts the ancient crystalline rocks from the younger sedimentaries. O. Nordenskiöld[51] elaborated the following classification:

A. Basin-shaped depressions on the continental shelf
B. Fjords in mountain regions
    I. Fold zones
        1. Radial fjords
            (*a*) Parallel with the strike
            (*b*) Transverse to the strike
            (*c*) In massive rocks
        2. Parallel fjords and fjord straits
    II. Finnmark type. Bedded and more or less horizontal rocks.

The Finnmark type is illustrated in Norway by the Porsanger, Laxe, Tana and Foranger Fjords and in Spitsbergen by Wijde Bay. It occurs in south Labrador.[52]

Distribution. Fjords are restricted to higher latitudes, the "fjord latitudes" or "drift latitudes" of Dana.[53] Thus there are no fjords in Africa or Australia (except Port Davey in south-west Tasmania[54]) or on the Asian mainland. The Red Sea and Persian Gulf, regarded by C. Ritter[55] as fjords, the inlets of Brittany, which E. Reclus[56] identified with fjords, and the submerged valleys around the Great Lakes which F. Ratzel[57] classified with them, are now by common consent excluded from the category. The Dalmatian inlets, Nordenskiöld's "Dalmatian type",[58] are also excluded

except by an occasional writer[59] who emphasises their U-form, hanging valleys and thresholds.

Penck's estimate[60] of 31,000 km for the length of the world's fjord coasts, like the figure quoted above (see p. 344), does not include those mapped during this century.

Fjords begin in Europe in 52° N. Lat. and in North America in 48° N. Lat. In the southern hemisphere, they appear in 43·5° S. Lat. in New Zealand and in 42° S. Lat. in Patagonia, i.e. 6–10° lower than in the northern hemisphere. Their distribution has been fully described.[61] They occur in Norway,[62] Scotland[63] (Loch Fyne, the longest, is 45 miles (*c.* 73 km) long; Loch Morar, the deepest, is 1017 ft (310 m) deep; and the length of the fjord coast is 250 miles or 400 km), Ireland,[64] Iceland,[65] Spitsbergen,[66] Novaya Zemlya,[67] Severnaya Zemlya,[68] Greenland,[69] British Columbia,[70] Alaska,[71] Labrador,[72] Newfoundland,[73] Patagonia,[74] New Zealand[75] (South Island), Kerguelen,[76] South Georgia[77] and the Antarctic[78] where they are more submerged and masked by ice than in South America.

Fjords are found mainly or exclusively on west coasts,[79] as in Scotland, Scandinavia, Iceland, New Zealand and Patagonia. In all cases they are confined to high lands. Where, as in Finnmark, for example, the high fells (*Högfjällen*) cease, the fjords become broader and shallower and less typical. This explains too why there are none on the north coasts of Asia and North America.[80]

Glacial origin. Although many older theories, such as those which postulated sudden catastrophes or erosion by the sea,[81] may now be safely omitted from our review, the problem of the fjord still awaits its complete solution. Since fjords pass upwards into typical fjord- or U-valleys the two are obviously homologous and had a common origin. Fjords may, therefore, be regarded as U-valleys on a vaster scale; their basins are comparable with rock-basins and their thresholds with rock-barriers. Submergence is an added complication. Yet the level of the sea is incidental, submergence controlling the fjord's physiography but not concerned with its origin. Epiglacial uplift converted many fjords in Greenland, Scotland and Norway into supramarine U-valleys[82] (see ch. XLV), or "freshwater fjords". Others termed "land-locked fjords"[83] are on the point of seceding from the ocean.

Consequently, it is not surprising to find the same conflict of opinion as was recorded in the previous chapter regarding U-valleys; some writers[84] regard the work of ice in connexion with their form and origin as negligible; others[85] ascribe them entirely or almost entirely to its erosion—ice removed thousands of cubic miles of rock in excavating the Alaskan fjords[86] and deepened the Sognefjord by *c.* 2500 m.[87] Glacial erosionists interpret the thresholds as solid rock, their opponents, in general, as moraines (see p. 350). Even the preservation hypothesis has been applied to the fjord basin.[88]

That glaciation has played an appreciable role in modelling the fjord is overwhelmingly clear and generally acknowledged. Most writers,[89] indeed, exclude from the category any that have not suffered glaciation. Since J. Esmark[90] first suggested that ice eroded the Norwegian fjords, but more particularly since Dana independently appealed to this agent for their excavation, the view has found innumerable adherents[91] and has been expressed for all fjord centres, including the Trondheimled[92] and extensions across the continental shelf.[93]

This theory alone explains adequately all the specified relationships. Thus

fjords are restricted to countries and latitudes which were formerly glacierised[94]—attempts to explain the distribution by appealing to climatic factors, such as the annual isotherm of 10°C or the polar limit of the subtropical rains,[95] must be deemed unsuccessful. They extend into lower latitudes in the southern hemisphere (see above) in harmony with its severer glaciation[96] and in Greenland are deepest where, in Scorseby Sound and Franz Josef Fjord, the ice was most expanded and its culmination nearest to the fjords.[97] Their U-shaped cross-section, hanging valleys, submerged basins, thresholds and greatest depths on the outside of bends,[98] features all diagnostic of glaciated U-valleys, are probably the best evidence. Their size is related to the glaciers which flowed through them so that neighbouring fjords vary enormously in depth—the Hardangerfjord is not so deep as the Sognefjord[99] and fjords in Iceland and the Faeroes are shallower than those of Norway or Scotland.[100] The walls are asymmetrical where the flow was faster on one side of the valley.[101]

These arguments are collectively decisive: they allow us to ignore those of a doubtful nature, e.g. the quantity of drift agrees with the erosion[102]; the deepest parts are where the tributaries joined the trunk glaciers[103]; or the glaciers would have calved at the head of the valleys had these been preglacial.[104]

Trough shoulders are rare because the preglacial valleys were very narrow. Troughs-in-troughs are also rare but have been seen occasionally.[105]

The preponderance of fjords on west coasts (see above) may be explained by higher elevation,[106] by harder rocks,[107] by meteorological factors, e.g. increased precipitation,[108] by more rapid flow, as in west Scotland, where the gradients were steeper and there was no opposing ice-sheet as on the east,[109] or (less convincingly) by a westward drag of the uplifted lands under the influence of the earth's rotation.[110]

The hypotheses already familiar to us but rejected in connexion with U-valleys and rock-basins have been applied to fjords too. For example, glaciers prevented the filling in of fjords[111] (see p. 265) and melt-waters, flowing sub- or extra-glacially[112] or laterally[113] (see p. 330), were responsible for the excavation. Interglacial streams, alternating with glaciation, have also been invoked[114] (see p. 331): H. Reusch,[115] in particular, has interpreted marginal ledges in the fjords, e.g. in the Sogne and Graven fjords, as unobliterated vestiges.

However much truth may lurk in these contentions, it is certain that the Glacial period seriously modified the relief of fjord lands and considerably deepened their valleys; the deepening has been computed at *c.* 600 m in British Columbia,[116] 1000 m in Norway[117] and 1100 m in the Sognefjord[118] (see above).

The glacier tongues undercut the sides of the marginal nunataks into steep, conical mountains or "back-fins" with a more or less circular base. The appelation given to such isolated cones, features of coastal mountains, fjord walls and islands, is *umáanaq* in Greenland[119] and *tinds* in Norway,[120] as in the Lofoten Islands and in the Syv Søstre (Seven Sisters), Torghatten and Hästman. Tinds may be completely isolated or may be simply more sharply accentuated points of the jagged lines of ridges. Unlike "horns" (see p. 306), they are rounded and marginal to the ice.

Adverse arguments. Apart from an unwillingness to ascribe such enormous results to glacial erosion, critics have opposed the theory of glacial

excavation because many fjords athwart the ice-flow are as deep as those which coincide with it[121]; the line of deepest water is sinuous and often wanders from the axis or to the outside of bends[122]; the fjords in Norway are too deep to have been filled with ice (the Sognefjord was occupied to a height of only 700 m[123]) or too wide to have been eroded by it[124]; they are absent from some areas of severe glaciation, as in east Scotland[125]; and they occur, as in Dalmatia (see above) or north Greenland,[126] in unglaciated regions or continue as depressions across the continental shelf.[127]

Modified rejuvenation gorges. These arguments, in so far as they are legitimate, are reconcilable with the theory already established for U-valleys, namely, that fjords are glacially modified immature valleys,[128] rejuvenated by terrestrial upheaval, less probably, by a lowering of the sea.[129] This view is prompted not only by the difficulties that the hypothesis of complete glacial excavation meets but for other reasons, several of them as applicable to fjords as to U-valleys. Thus the shoulder is rounded[130] "like a human forehead" as Richter[131] expressed it; the *Saekkedal* or blind and steep inner end is the analogue of the *Talschluss*[132]; the fjord passes into the typical U-valley, the "fjord valley", now generally recognised as a modified rejuvenated valley; remnants of preglacial forms and small V-shaped lateral valleys are preserved in places[133]; and the valleys, e.g. in British Columbia and south Alaska, which traverse the width of the coastal ranges, are of high antiquity.[134]

The elevated and dissected peneplains which sweep back from the top of the fjord walls in Labrador,[135] Spitsbergen,[136] Iceland,[137] Greenland,[138] Norway[139] and New Zealand (see below) are probably the best vindication: they demonstrate a late-Tertiary rejuvenation (occasionally denied[140]). This plain is best known in Norway; it is the *fjeld, fjell* or *vidde*, the "palaeic plain" of Reusch.[141] Surmounted by monadnocks, e.g. Jotunheim and Dovre and itself subject to a *Grossfaltung* (Norw. *Storfoldning*) in the opinion of some geologists, it stretches above the chasm-like fjords as a broad, open and gently undulating platform, studded with innumerable lakes and marshes and drained by wide and mature valleys. Its highly ice-worn surface, which abounds in roches moutonnées and is bestrewn with erratics, supports only a scanty vegetation of moss, lichen and a few shrubs. Though influenced by the hardness of the rocks, it is mainly independent of both strata and structure. Reusch recognised two peneplains, a higher one of Mesozoic age at 1400–2000 m and a lower, Tertiary one between 600 m and 1400 m. Other morphologists[142] also recognise them —their age has been given as Miocene and early Pliocene[143]—though the palaeic surface is also thought to be an overthrust plane or a re-appearing Pre-Cambrian surface.[144] In south Norway and in north Scandinavia three or more plains are developed,[145] in general with vertical intervals of 200 m.

A similar uplift elevated the "intermediate peneplain" of Scotland[146] which occurs at 800–1000 ft (*c.* 240–300 m) and succeeded the intrusion of the Tertiary dykes and Arran granite: the "fjordland peneplain" of middle Tertiary age in New Zealand[147] is its analogue.

Tectonic influence. Valleys and fjords gave rise to a keen controversy in the history of Geology which ended over a generation ago in the temporary victory of the erosionists over the advocates of the outworn tectonic hypothesis (see p. 315). The connexion of fjords with tectonic lines, perhaps first stressed by R. I. Murchison,[148] has often been demanded[149] to explain the

behaviour and form of fjords, especially the *Spaltenfjorde*. Fjords, it is said, are generally restricted to uplifted lands and coincide with longitudinal and transverse coasts[150]; they are straight and lie along faults or joints,[151] as in north Helgeland, in the Hardangerfjord and in Oslofjord whose deepest part is tectonically most depressed; their plan resembles in its straight and parallel lines, e.g. in islands, peninsulas, bays, bends and angular intersections, a system of fractures rather than a dendritically branching river-system[152]; fjords are parallel in the Julianehaab granite with the most important dykes radiating from the batholites[153]; and they dissect folded regions in north Greenland.[154] The closeness and in some cases insignificance of their drainage areas are inconsistent with river-channels.[155]

Gregory,[156] who repeatedly urged the connexion, recognised the following three types in Scotland; (*a*) rift valleys, e.g. the Sounds of Sleat and Islay; (*b*) shatter-belts, e.g. Kerrara Sound; and (*c*) subsidences along single faults, e.g. Lochs Linnhe, Lomond and Maree. O. Holtedahl[157] has repeatedly emphasised the importance of marginal fault lines in the shelf areas off some high northern lands, e.g. Scandinavia, Spitsbergen and Greenland, though ice has modified the river-channels into trough-like basins. Drygalski[158] related fjords to contraction processes in polar regions.

The view that fjords are preglacial valleys coincident with tectonic lines, e.g. axes of folds, shatter-belts, crush-zones, joints and contacts of geological formations, has been widely expressed for most fjord countries[159] (Scotland, Norway, Iceland, Faeroes, Spitsbergen, Greenland, Labrador, British Columbia, New Zealand and Kerguelen). Faults control the features in several parts of Fennoscandia,[160] and in Norway the fjords run along the axes of folds or of fissures normal to them while in the stiff foreland which lacks Caledonian lines they are small and rare.[161] The Sogne and Hardanger fjords lie in synclines of relatively weak sediments or schists enclosed by massive crystalline rocks.

This relationship is undeniable: yet geologists differ concerning the extent of such influence. Some regard fjords as purely tectonic,[162] open chasms, slightly widened by clearing out crushed and broken rock (the parabolic cross-sections in the west Greenland gneisses were due to the curvature of the joints[163] and the Morkefjord, east Greenland, was younger than glaciation[164]). Others think that ice-erosion was only secondary and selective.[165] Others again consider that tectonic lines merely guided and facilitated denudation.[166]

The reaction against the catastrophic school of Geology probably went too far in its virtual neglect of the tectonic influence upon valley direction and form. Tectonic lines, with crush zones, infaulted strips and weak rocks, have without question given many fjords their rectangular and geometrical pattern (they nullify the objection that such rejuvenated valleys should have incised meanders[167]). Yet tectonic forces did not cause the openings; they merely provided lines of weakness favourable to rapid erosion by water and ice.

Such movements are probably more important in the case of the Norwegian Channel which, averaging 104 km in width, skirts south Norway for 890 km and descends from a threshold at −273 m to −382 fathoms (699 m) near its head[168] (see fig. 232). Although it may have been the course of a preglacial river,[169] possibly the *Alnarpsflod*[170] (see p. 281), and have been moulded somewhat by ice[171] (cf. p. 212), it resulted primarily from late-Tertiary trough faulting[172] or flexuring.[173] It is related to definite tectonic lines in Scania

and central Sweden, to dislocations which skirt the Norwegian coast from north to south[174] and to present seismic disturbances.[175]

Nature of thresholds. The most important factor in the fjord problem is the nature of the thresholds. On the analogy of the fjord-valley, these are probably not all alike; some may be moraines, others solid rock with or without a morainic cover. Soundings alone fail to distinguish them and conclusive proof is in any one case hard to obtain. They are ascribed to morainic material landslipped from the sides of the valley[176] or to alluvial bars,[177] as has been established for example for the mouth of Yakutat Bay, Alaska, and for some Patagonian fjords. They are also regarded, either generally[178] or in particular regions,[179] e.g. Iceland, Scandinavia, Alaska or Patagonia, as terminal moraines piled up on a floor which sloped outwards with a normal river profile. The valleys, it is said, continue across the continental shelf[180] but are obscured by sediments, chiefly drift; the shoaling of the Norwegian Channel opposite the coast between Bergen and Stavanger has been so explained.[181]

Some thresholds are indubitably partly moraine.[182] Yet occasional erratics dredged from a threshold, as in Greenland,[183] do not prove it to be morainic any more than waters flowing over rock prove it to be solid throughout. Nevertheless, many geologists[184] urge that the thresholds are solid, including that of the Sognefjord[185] which is overlain by 158 m of water and is made of hard Devonian conglomerate. Rock is exposed[186] in them or in artificial sections or in a series of islands which spans the fjord, or, if just awash as in Iceland or at Connell Ferry, Loch Etive, Scotland, gives rise to falls at low tide. Moreover, thresholds occur where on the glacial theory they should be expected,[187] e.g. where the valley changes its character or where the ice-pressure was diminished by spreading, by the obstacle of the skerry guard, or by melting. They are associated with spurs and contractions at the entrance to sounds and are analogous with barriers in fjord-valleys. The enormous moraines that the rival hypothesis requires are highly unlikely and unknown elsewhere. They conflict in shape with the extremely low slope of the *contre-pente* (see p. 343).

The basins behind the thresholds are of course an integral part of the question. They have been attributed to ice-erosion[188] because they resemble the rock-basins of U-valleys; their depths vary in adjacent fjords (basins and swells are the expression of a rhythmic plunging and rising of the glacier related to incoming tributaries[189]) and in Finnmark coincide with gneiss-granite outcrops which facilitated plucking[190]; and the deepest Norwegian fjords, e.g. Sogne and Tys fjords, are situated in uplands which were sufficiently big and high to ensure a considerable flow.[191] Although this gives 1100 m for the glacial overdeepening of the Sognefjord (see p. 347) erosion of live rock on this scale is repudiated by those who would restrict it to detraction[192] (see p. 271).

Several geologists,[193] following A. E. Törnebohm,[194] imagine a backward tilting of the country like that postulated for the subalpine lakes (see p. 278), especially those of north Italy to which, as Peschel pointed out, the basins present a striking morphological likeness. De Geer[195] stressed this origin and attributed fjords to a fracturing which accompanied a late-Tertiary upheaval of the continental borders around a sinking "Atlantik" and "Skandik" and a radial displacement of the magmas. The extension to a force acting all round the North Atlantic and as far as the Alps[196] is im-

probable. However plausible De Geer's hypothesis may be, it is very doubtful whether a tilting of the required complexity can be accepted as a general cause. Indeed, the strandflat's horizontality and the behaviour of the older peneplains show that it has not taken place.[197]

In fine, the greater depth of the inner end of the fjords has been explained by ice-erosion—the action began here and lasted longest[198]; by plucking in fissure-zones[199]; by bending and tectonic movements,[200] possibly still continuing in north-east Greenland[201]—the greatest depth lies in the area of deepest *Spaltenbildung*[202]; by deposition of moraines and drift at the mouths of preglacial valleys[203]; and by backward tilting due to tectonic movements or isostatic depression (see below).

Submergence of fjords. Since Dana's paper of 1849, the opinion has been widely held[204] in both Europe and North America that fjords are preglacial stream channels modified by ice and drowned by an epiglacial submergence from which the fjord countries have only partially recovered[205] (see ch. XLV). Submergence (which is naturally more difficult to prove than emergence) is demonstrated, it is said, by submerged hanging valleys, e.g. in College Fjord, Alaska, and in British Columbia; by submerged confluence steps,[206] cirques and pot-holes, as about Saltenfjord, Norway[207]; by valleys in the continental shelf (see p. 1240); and, less convincingly, by submerged coasts south of the glaciation, as in California.[208] But the strandflat lies near present sea-level (see p. 1250) and freshwater deposits of interglacial age at Hernösand, Norway, occur also just at that level.[209] Moreover, the continental shelf persists at the same depth around west Europe; and the world's shores outside the glaciated regions have not been similarly submerged. Consequently, as noticed already, backward tilting of the fjord-lands is substituted for direct submergence. The cause of the tilting is obscure; it is sought in an isostatic depression, a big part of which has persisted[210] (see p. 1327), or in some action possibly of a tectonic or deep-seated nature.[211] Marginal flexuring has also been invoked.[212]

Glacial erosionists, almost until the end of the last century, believed that a late submergence drowned the valleys. The necessity for it rested upon the assumption, usually made and occasionally given quantitative form, that when a glacier entered a sea too shallow to float it, part of the ice equal in weight to the displaced water was sustained so that the glacier's pressure on its bed and its erosive efficiency were correspondingly diminished. Gilbert,[213] following a lead by others,[214] disputed this. He showed that the basal erosion remained unreduced so long as a glacier kept contact with its bed and there was no film of sea-water along its sole. The only possible reduction arose from the slowing down of the glacier by the mass of water to be displaced, and even this may have been more than counterbalanced by the increased flow due to calving at the snout (see p. 104)—a glacier 900 m thick would continue to erode its floor even when this was submerged to a depth of *c.* 820 m.[215] Gilbert's view was supported on general grounds[216] (W. M. Davis[217] found evidence of sublacustrine glacial erosion in Montana) and applied to New Zealand[218] and South Georgia.[219] Thus fjords were glacially eroded to their present depth when the land bore approximately the same relationship to sea-level as it does to-day.[220]

Conclusions. Fjords were probably originally deep and wide canyons with short spurs and hanging valleys resulting from Tertiary rejuvenation.

Their courses correspond with the regional trend and slope of the country, with the variable resistance of the rocks and structures, and with the prevalent joint and fault systems; their depth and character were determined by running water and weathering. Ice-erosion along these valleys, though absolutely great, was relatively small,[221] except in the deep fjord-basins. The spurs were faceted or removed, the sides straightened and smoothed, the valley-heads steepened and exaggerated, and the rock-basins hollowed into their floor. The present submergence is incidental in the sense that the height of the floor depends upon erosion: the ice retired and the sea came in (pl. XIB, opposite).

2. *Fjärds*

Form and distribution. Fjärds are more or less parallel valleys which exhibit curved lines, gentle slopes and indented and relatively low, island-strewn shores.[222] They differ, therefore, from fjords in their broader and

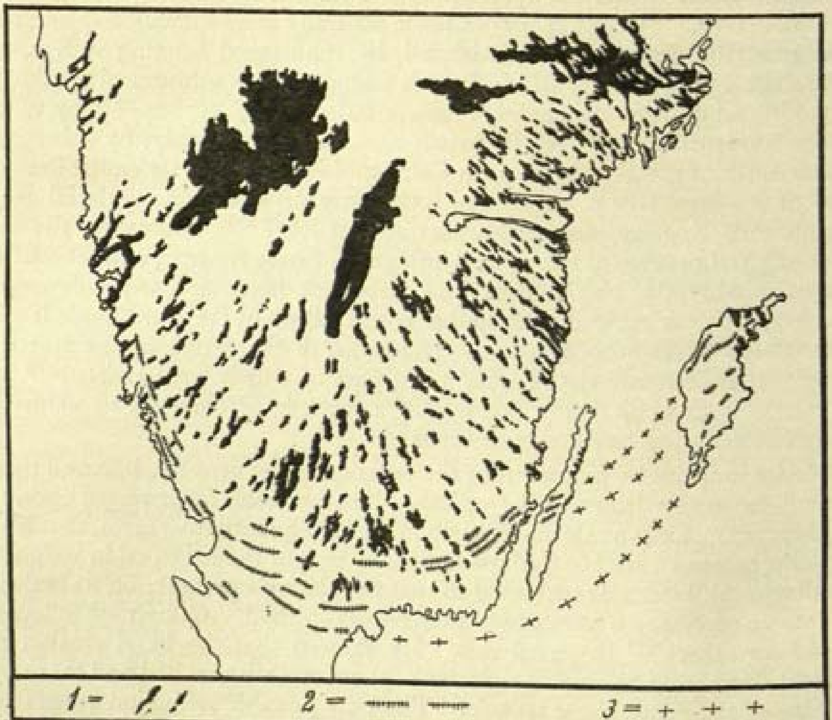

FIG. 64.—Map of the fjärds and radiating system of lakes and winding valleys of south Sweden. 1, Lakes and fjärds; 2, End-moraines (diagrammatic); 3, Ice-margins (hypothetical). E. Werth, *Z. Gl.* 8, 1914, p. 345.

less regular outlines[223] and from *rias* because their wider parts frequently include deep basins, sometimes with rock-bars and thresholds.

Fjärds are found in Sweden between Warberg and Oslofjord where they change into the fjord type.[224] They are also seen on the islands of Öland and Ösel and in Estonia[225]; in south Norway[226]; in Scotland[227], Loch Crinan, around Arisaig, in the Shetland Islands, Barra and Lewis; and in Nova Scotia, Newfoundland and parts of north Canada.[228] The inlets of

A. Schliffgrenze and Schliffbord, Grimsel Pass
[Transmitted by H. Louis]

B. Glaciated scenery, Ikeragsuqsuak, Greenland, with U-valley and a fjord showing the " accident " of submergence [A. Courtauld]

A. Section in boulder-clay near Killiecrankie station, Perthshire, Scotland
[Geol. Surv. Gt. Britain : Crown copyright]

B. Drumlin country looking north-east over Castlecaulfield, Co. Tyrone
[J. K. St. Joseph : Crown copyright]

Maine, U.S.A.[229] belong here with the exception of Somes Sound, Mount Desert Island, which is a fjord. Kerguelen has transitional forms.[230]

Origin. Fjärds, unlike fjords, occur in lowlands or peneplains. They are restricted to glaciated areas and run parallel with the ice-flow and, according to De Geer,[231] with joint systems. Werth,[232] who discussed their relationship to fjords and föhrdes, regarded fjords as the product of valley glaciers, fjärds of ice-sheets. He considered that the south Swedish fjärds continued in the periphery the great radial system of lakes and winding valleys of the interior (fig. 64): the lake chains and radial basins were parallel with the ice-flow and were formed, in intimate connexion with the osar, by radiating subglacial streams flowing eastwards and upwards towards the edge of the ice and the least pressure. At the centre of the fan lie the big lakes of Sweden. Gregory[233] ascribed them to a partial submergence of a low, irregular surface of hard rocks while P. Stolpe[234] emphasised that the rocks in fjord and fjärd countries were different. Fjärds, therefore, are preglacial valleys in lowland relief modified by subglacial waters and direct ice-erosion.

3. *Föhrdes*

Form and distribution. Fordes or Föhrdes (*Föhrden*), which Gregory[235] fully described, are relatively shallow, submerged valleys which branch at their heads into several subaerial valleys, containing a number of lakes separated by shallows. The gradients of their floors vary from 1 in 300 to 1 in 1400.[236] Flensburger Föhrde, 40 km in length, is typical[237] (fig. 65).

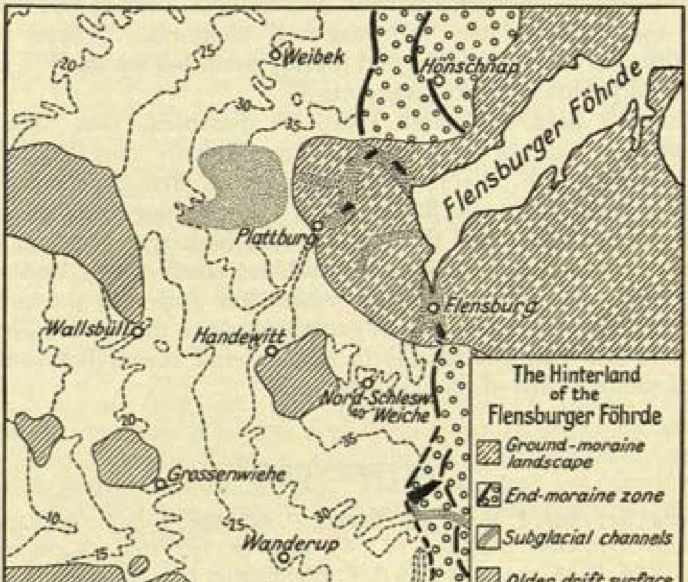

FIG. 65.—Flensburger Föhrde. P. Woldstedt, *J. L.A.* f. 1921, p. 794, fig. 4.

Föhrdes are found in Schleswig-Holstein and Denmark between Kiel Bay and Limfjord and Mariager Fjord—they were formed in association with the Dalecarlian–Baltic ice[238] and constitute a Cimbrian coastal type. They also occur on the coast of north Canada and the offlying islands[239] on Long Island, New York,[240] and in north-west Patagonia.[241] There are none on the southern Baltic coast because the surface forms are different[242] or current-borne detritus has converted them into Haffs.[243]

Werth[244] maintained that there was no genetic difference between fjärds and föhrdes. Dinse,[245] however, affirmed that föhrdes characterise low, young coasts, as in the Cimbrian Peninsula, and fjärds low, old coasts. Drygalski[246] distinguished between fjärds in regions sloping away in the direction of ice-flow, as in Sweden, and föhrdes in lands, such as Holstein, which slope towards the ice.

Origin. Föhrdes have occasionally been deemed to be subaerial valleys which have been tilted and submerged.[247] More commonly, they have been considered to be drowned *Rinnenseen* (see p. 241) or *Tunneldale*,[248] the higher parts, as in Schleswig,[249] occupied by freshwater *Rinnenseen*. Subglacial (or interglacial[250]) streams eroded the valleys[251] (possibly along tectonic lines[252]). Thus cones and dejection deltas were deposited at their mouths and they were connected with osar: if the ice changed its direction new tunnel valleys may cross an older set of tunnel valleys[253] (cf. p. 421). Ice-erosion has sometimes modified them during periods of oscillation,[254] as by K. Olbricht's Baltic readvance,[255] to form his "exaration landscape". Some, indeed, ascribe the basins solely or essentially to ice-erosion[256] and regard them as *Zungenbecken* (see p. 270): the valleys narrow and shallow towards their distal end and their shape is not that of subglacial channels. Others prefer to think the föhrdes were preglacial and merely moulded by ice.[257] The bars have been interpreted as retreat moraines.[258]

It is generally held,[259] though Werth[260] dissented, that their present drowned state is due to later submergence, usually correlated with the Littorina period.[261] F. Wahnschaffe[262] observed that they are bound up with the lakes, lake-chains and *Rinnenseen* of the irregular moranic accumulations south of the Baltic. If the moraines are high and fairly far from the coast, as in Pomerania, Mecklenburg and the south Baltic region,[263] the hollows are occupied by fresh water; where low and near the coast, as in Schleswig and Denmark, the sea has gained access to them.[264] The *Bodden* of Pomerania and east Mecklenburg, shallow and round hollows, are the submerged round basins of the lake-plateau.[265]

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(*B*) *GLACIAL DEPOSITION*

CHAPTER XVI

DRIFT

Importance. Drift, a term introduced by C. Lyell[1] and widely used to-day though without the early implication that it was "drifted" by water or floating ice (see ch. XXX), is a generic name for accumulations referable directly or indirectly to the Pleistocene ice. It comprises much of the loose and incoherent superficial deposits, such as clays, sands and gravels, in glaciated countries. Unobtrusive in appearance—unlike glacial erosive features it is rarely scenically impressive—it seldom rivals in thickness the older geological formations, notwithstanding the exceptional figures occasionally recorded (see p. 223). Nevertheless, in its distribution, the drift is scarcely, if at all, inferior to preceding formations; its extent has been estimated at 7–8% of the land-surface of the earth[2] (36% of Europe, 23% of North America) and its volume at 600,000 cu. km.[3]

A study of the drift is necessary in connexion with most constructional works, e.g. for canals and roads, for surface and underground railways, and for foundations of all kinds. The drift is an important source of underground water (obtained from fluvioglacial gravels, terraces, loess and interglacial horizons), as in Russia,[4] and of materials for buildings, roads, railways, dams and canals, of sands for cements, glass-making, moulding and filter purposes, of clays for bricks and tiles and of diatomite, for various uses; their gainful utilisation calls for all-round practical and theoretical studies. The materials are being removed on a very large scale. Thus a year's consumption of glacial gravel in Sweden at present amounts to about 7 million cu. m, representing a length of about 35 km of an ose of normal size.[5]

Glacial or glacigene[6] deposits are also of prime importance to the agriculturist. To them the land owes much of its fertility and even much of its ability to be cultivated: in hilly areas, for example, the drift boundary often marks with great precision the limit of tillage. The drift is on an average very much more fertile than the older rocks, since detritus derived mechanically from many kinds of rock containing a high percentage of soluble plant foods, has led to a wide diversity of soil. Glaciated Wisconsin, for example, is much superior to the unglaciated part of the state in the area, value and productivity of the farmlands and in the density of the rural population.[7] In England Anglo-Saxon settlements are related to the gravel distributions.[8] The various types of drift have their influence on the distribution of the water plants in north Europe.[9] The effect was only detrimental in mountainous regions where clearing away the stones has been one of the hardest tasks of the cultivator.

General character. The drift is exposed in artificial sections and in raw scars of streams and coasts and in runnels and gashes made by rain. Its depth, which may introduce magnetic anomalies,[10] varies much according to the underlying relief, to the amount of loose debris strewn in the path of

the ice, to the changing conditions under and at the margin of the ice, and to the resistance of the local rocks to glacial erosion. It is unusually great where, as in north Germany, soft rocks (Cretaceous and Tertiary) were readily converted into drift. Thick drifts are commonly found in valley bottoms, across the mouths of tributary valleys, in the lee of hill spurs, and in places where the ice had opportunity to spread out beyond a pass or narrow valley which had restricted its flow.[11]

The thickness, if there are enough data, can be portrayed by "isopachytes" or lines of equal thickness. Such maps, which are readily constructed,[12] have been drawn for many areas, including the Polish plain,[13] the morainic terrain of south-west Finland[14] and the districts around København[15] and Königsberg.[16]

Glacial deposits were produced mechanically and chemically. Even small fragments in coarse sands are angular, broken and unabraded; their shape is usually governed by rock-structures.

Classification. The tendency of strata to be impersistent and less consolidated and to alternate, which is increasingly revealed as we proceed through later geological history, culminated in the Quaternary era. The rapid change in facies and local extent of the beds is carried further in this era; "Diluvium is chaos".[17]

The drift's topographical expression is as varied as its structure and arrangement. Studies in glacierised and glaciated areas, just as in the case of glacial erosion, supply keys to the glaciological processes which made the drift. All types of glacier, however, need to be examined, since the ice in its waxing and waning passed through all of them. Too rigid adherence to lessons learned in the Alps has led to misconceptions and errors. Even modern ice-sheets which best exemplify the Pleistocene conditions are unfortunately of little help; they either end in the sea to hide their marginal phenomena from inspection, as in the Antarctic, or they terminate on lands of hard rock which generate little detritus. Thus there are almost no moraines in east Greenland,[18] and fluvioglacial debris on Jameson Land, north Greenland, has only a small distribution and an origin which is by no means certain.[19] Moraines may, however, occur on the fjord floors.

The infinite diversity of the drift makes diverse interpretations possible and inevitably introduces the individual judgment of the investigator in classifying them. Several attempts have been made to do this[20]: the basis may be structure, form, genesis, or time-relation. Form, the basis of W. J. McGee's scheme,[21] is like structure alone insufficient. Time-relation would doubtless be useful were our knowledge not so imperfect. J. B. Woodworth[22] has employed the relation of the ice-contact. Genesis, combined with form and structure, is the ideal basis and underlies to some extent the methods of G. De Geer[23] and T. C. Chamberlin.[24] The subjoined classification, which takes these conditions into account and is largely based on Chamberlin, recognises the main types which in the past have been differentiated. It is necessarily imperfect since the genesis of many glacial accumulations is still obscure.

The ill-defined nature of many patches of stratified drift, in particular the incipient valley-trains, outwash fans, kames and osar, makes it very often impossible to assign to them their proper place in the classification with any degree of confidence. It but emphasises the complexity of the processes of deglaciation which the diverse accumulations of the table suggest.

Classification of the Drift

I. Within the Area of the Ice
- 1. By en- or subglacial streams
  - (i) Osar (in part)
- 2. By ice and streams
  - (i) Erratics
- 3. Ground products of ice
  - (i) Boulder-clay (*moraine profonde*)
    - (*a*) Till sheet
    - (*b*) Drumlin
    - (*c*) Crag-tail
    - (*d*) Pre-crag
  - (ii) Marginal moraine
    - (*a*) Lateral
    - (*b*) Terminal

II. Extraglacially—fluvioglacial
- 1. Free drainage
  - (*a*) Kame-moraine
  - (*b*) Osar (in part)
  - (*c*) Fan
  - (*d*) Valley train
  - (*e*) Outwash apron
  - (*f*) Outwash and pitted plain
- 2. Ponded drainage (shallow or deep)
  - (i) Glacio-lacustrine
    - (*a*) Delta
    - (*b*) Terrace or beach
    - (*c*) Lake-warp
    - (*d*) Ice-rafted erratics
  - (ii) Glacio-marine (shallow or deep)
    - (*a*) Delta
    - (*b*) Glacio-natant boulder-clay

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CHAPTER XVII

ERRATICS

General. Erratics are fragments of rock which ice or its streams have transported from their parent source. The limitation to fragments which had travelled only an appreciable distance ceased to be imposed within a decade of its suggestion by J. de Charpentier: the distance is limited only by that the ice travelled.

Erratics, which may be embedded in till or rest on rock, on other erratics or on any kind of drift, vary in size from small pieces to huge boulders. The latter, the erratic blocks[1] of Brongniart and De la Beche, early attracted attention because of their dimensions and often impressive difference in composition and colour from the underlying rocks. They were topics of popular legend as in Scotland[2] where they served for the purposes of commemoration, congregation, worship, alignment, sepulture, the trial of offenders, the contracting of engagements and as objects of superstition. They were used in prehistoric monuments, as in standing stones, dolmens and stone circles, especially where they abounded, e.g. in Sjaelland, Holland and north Germany,[3] and in later monuments, as in gravestones, the pedestal of Peter the Great in Leningrad,[4] or the Hindenburgstein of Tannenberg[5] (Rapakivi granite from the Åland Islands).

In early days, the larger ones at least were preserved: they were venerated in connexion with cults or religions. In later times, though some, as the result of the efforts of A. Favre in Switzerland and A. Falsan and E. Chantre in France, have been declared inviolable,[6] they have generally been ruthlessly destroyed or removed through the activities of the agriculturist, builder and engineer. They have been used to pave and wall roads and to build castles and churches, as in East Anglia or Russia. Erratics of Scandinavian granite have been employed in north German villages[7] and in dykes and harbour walls along the Dutch and German coast,[8] and limestones have been burnt for lime. Roads in Germany, Poland and Russia since time immemorial have been paved with them—in Mecklenburg-Schwerin over 1,650,000 cu. m of erratics[9] have been so used. Their abundance in mountainous areas impedes agricultural operations and here and elsewhere has caused them to be used to border fields.

Erratics have enlarged our palaeontological knowledge of particular formations[10] as in the case of Bryozoa, Ostracoda and Phyllocardiae, the trilobites and brachiopods of Prussia, or the molluscs of the Senonian. They have helped in tracing valuable ores, as in Fennoscandia[11]—D. Tilas noticed this in 1740—and of gold, tungsten and coal in Nova Scotia,[12] as well as in ascertaining the composition of parts of the inaccessible sea-floor (see p. 369) or of regions beneath the drift, as is demonstrated by the *einheimische Geschiebe* of the North German Plain,[13] including the facies of a particular horizon, e.g. the chalk of the Baltic.[14] They prove, for example, that rocks still undiscovered *in situ* occur in Scandinavia,[15] that Pliocene lies beneath north Germany,[16] and that Bracklesham beds formerly existed north of Finchley in the London Basin.[17] They have revealed the secret of the "blue quartz"

whose colour is due to almost ultra-microscopic needles of rutile.[18] So important are they that a special study is devoted to them: the *Zeitschrift für Geschiebeforschung* (1925–42) provided for the publication of the results. The amber used by palaeolithic man in the Moravian Gate may have been transported thither from the Baltic region by ice.[19]

The most resistant and durable materials have usually travelled farthest and survived the journey best. Hard and well-jointed rocks, e.g. from dykes or sills, are generally well represented and granites have a disproportionately large share.[20] They all, however, become smaller, rounder and fewer as they move away from their source—G. B. Greenlough[21] discovered this as early as 1814 and Murchison[22] noticed it on the plains of Russia. Small pebbles in particular disappear rapidly, resistance being inversely proportional to size.[23] Rocks, like the Skiddaw Slates of the Lake District,[24] which do not lend themselves to plucking have few erratics.

Gigantic erratics. Erratics are sometimes gigantic. Classical examples are the boulders, accommodating quarries, near Monthey, and the *pierre-à-bot*, of protogine granite from the Mont Blanc massif and of 1370 cu. m content, on the flanks of the Jura Mountains 100 m above Neuchâtel.[25] Switzerland has numerous other examples.[26] Scotland has many erratics each over 100 tons in weight.[27]. Others, usually of Secondary strata, occur elsewhere in Britain, notably near the North Sea from Caithness to the eastern counties of England where they are mostly in the Great Chalky Boulder-clay. Mention may be made of the Leavad erratic of Cretaceous rock, *c.* 200 m by 137 m by 8 m, in Caithness[28]; the Coniston boulder, 46 m long, near Edinburgh[29]; the Kidlaw limestone boulder of East Lothian, *c.* 530 m by *c.* 400 m[30]; limestone boulders extensively quarried for years in Kilmux Den, Fife[31]; a "raft" of Great Limestone, fully 800 m long, in Northumberland[32]; Lincolnshire Limestone at Great Potton and other places, over 300 m long[33]; Lincolnshire Oolite at Marston,[34] 275 m by 90 m north-west of Melton Mowbray,[35] and 180 m long at Beacon Hill, Rutland[36]; Gault, Cambridge Greensand and Chalk, 412 m long, in Kimmeridge Clay at Roslyn Hole near Ely[37]; Chalk at Catsworth, Huntingdonshire, with a village perched upon its back[38]; Ampthill Clay, 20·5 m thick, at Biggleswade, Bedfordshire[39]; and Kimmeridge Clay above Greensand near South Runcton, Norfolk.[40] Elsewhere in Britain large boulders are scarce; they include a Carboniferous Limestone erratic, 90 m long, in Anglesey[41] and a grit boulder, 180 m long, near Abergavenny.[42]

Comparable masses have been described from the mainland of Europe, e.g. near the Baltic, near Leningrad and the Valdai Hills, and north-west of Moscow in Russia[43] (farther south they broke up in transport); in Finland where they are very common and as big as cottages[44]; Chalk erratics in Scania, up to 850 m long and 15 m thick[45]; and similar Chalk erratics,[46] several miles or kilometres long, in Sjaelland, Schleswig-Holstein, Mecklenburg and Pomerania.

Germany's biggest surface erratic is the *Grosse Stein*[47] of Gr. Tychow in Pomerania, a garnetiferous gneiss 16·9 m long, 11·25 m broad and 44 m circumference. Denmark's *Damesten* is a granite mass 45·8 m in circumference.[48] These are, however, far surpassed by the huge masses (*Schollen*) which have been frequently described and figured.[49] The largest are up to 4 km by 2 km and 120 m thick.[50] About one-half are of Tertiary and one-third of Cretaceous clays and sands. The rest consist of interglacial

(especially in Schleswig-Holstein and Denmark) and glacial deposits with an occasional mass of Palaeozoic rock.[51] These *Schollen*, about 450 in number, are almost all in the Upper Diluvium (only in Schleswig-Holstein are they plentiful outside the later drift) and rapidly become fewer to the south. Gravity and magnetic anomalies are related to their distribution.[52]

Similar erratics are not unknown in North America. Limestones, weighing 13,500 tons, in Ohio a few miles from their source,[53] thin Ordovician shale, acres in extent, in Minnesota,[54] the Madison Boulder (granite) near Conway, New Hampshire[55] (10,000 tons), the "haystacks" of basalt, weighing thousands of tons each above the Columbia valley,[56] and a quartzite erratic in south-west Alberta, measuring 24·5 m by 12 m and 9 m high and weighing 18,150 tons,[57] may serve as examples.

Method of dislodgment. The hugeness of these erratics led early writers, who only imperfectly realised the transporting power of ice, to consider them to be *in situ*, as in Scania,[58] or to have faulted junctions, incredibly complex, as about the Roslyn Boulder.[59] Interest is now focused upon the actual way the ice quarried them, notably those which are thin compared with their superficies. That the method varied is certain. A few slipped from rocks on to the ice[60] or originated when drift-filled roofs of subglacial stream tunnels collapsed.[61] Many were removed from steep coasts (e.g. the Glint) or valley sides[62] or from the lee faces of cliffs or ridges—the gigantic erractics of Marlstone which trail south of Grantham for 12 miles (*c.* 20 km) south of east were plucked from the disturbed and broken escarpment near that place.[63] Ledges were pressed into slight folds without fracture or transportation, or were disrupted into erratics[64] covering acres of ground or measuring 30 m in length. Outliers were dislodged in this way.[65] In a boring 280 m deep near Rostock, Tertiary rocks and drift were repeated six times.[66]

To secure the irregular surface thought necessary for such quarrying,[67] as well as to explain the structures of Möen, Rügen, etc. (see p. 257), German geologists have invoked interglacial fault-scarps,[68] asserting that the *Schollen* are distributed along north-west–south-east tectonic lines[69] or that their thickness is related to the throw of the fault.[70] This derivation, though supported by the sedimentary petrography of the drifts,[71] is questioned by those who ascribe the disturbances to glacio-static forces (during the Warthe or Brandenburg advances[72]) or derive the erratics from the cliffs[73] of interglacial valleys or the sea or from the action of frost and plucking.[74]

Other large erratics of plastic material have been sheared off the summits of folds produced by advancing ice (see p. 256) or obtained by thrusting in frozen ground,[75] the sand, clay, etc., being carried as frozen masses or "frigites".[76] Many were prized off by tongues of boulder-clay, sand, etc., squeezed along bedding planes in strata beneath the ice, especially on rising ground or in the lee of deep valleys athwart the flow,[77] perhaps during a sudden advance following a long pause.[78] The beds broke along joint planes, opened possibly by frost (see p. 301), and moved forward bodily. Such intrusive wedges have been mentioned many times since J. Hall[79] first recorded them in New York in 1843, e.g. in North America[80] and in Britain[81] (in Trias in Cheshire and Elgin—the drift penetrated 360 ft (110 m) from the face—in London Clay at Upminster, in thin Crag sands in East Anglia, and in Chalk and Mesozoic clays). They have been found also in Sweden,[82] as by O. Gumaelius[83] who appreciated their significance, and not infrequently in north Germany where beds of brown coal are mined in the drifts of Saxony.

The Glacier de Trient in 1818 insinuated itself between the solid rock and soil.[84] If the strata dip icewards, the blocks may not only be twisted out of position but tilted at a high angle or inverted, like an arm still united at the base, and pushed forward out of place.[85]

Erratic-trains. Erratics lie on or within the drift in lines extending from the lee of crags or knobs of igneous or other rock. To furnish such a stream, the *trainée* of early geologists, a parent rock should satisfy certain conditions; it should be hard, suitable for plucking, sufficiently distinctive to be readily identified ("kenspeckle" rocks[86]) and have a limited and well-delineated outcrop. An example of a linear train is the Snake Butte train in Montana.[87]

All great dispersals and most small ones are more or less fan-shaped, as Agassiz[88] was one of the first to realise. The comet-like train has an axis which coincides roughly with the local ice-flow and a terminal width much bigger than the outcrop, both horizontally and vertically. The fanning may have been caused by subglacial streams[89] or marginal melt-waters[90]; by drift-ice,[91] especially during earlier and later stages, as in the Yoldia Sea; by radiating flow near the ice-edge,[92] e.g. in the Oder Lobe; or by varying movements during the opening, peak and closing stages, as oblique valleys and topographical irregularities affected the ice[93] (see p. 1146). It also resulted when the ice radiated from a centre, as in the case of the granite boulders from Barnesmore, Co. Donegal[94] (fig. 143, p. 780), and was most pronounced if the flows diverged through direct impact of ice-masses (see p. 715), as exemplified by erratics of the nepheline syenite of the Kola Peninsula which were conveyed to the north and south-east and across the peninsula between.[95]

Erratics within a fan rarely travelled by the shortest route but according to chaotic intercrossing lines of successive times.[96] Subglacial and extraglacial streams, wandering centres of radiation[97] (see p. 670), and incorporation from earlier drifts or fluvioglacial or interglacial deposits varied the boulder content vertically and crossed the trains—erratics during the last glaciation in Denmark came first from the north (south Norway and west Sweden) and later from the north-east (east Sweden and the Baltic). They confused the "homochrone" erratics,[98] i.e. those distributed at any one time, like those in the upper drifts of certain areas of Finland.[99]

The boulders on the margin are sporadic, so that a fan's precise limits are not easy to define, but become more and more numerous towards the axis, though local relief, e.g. the Estonian Glint (= cliff), introduces variations by imposing obstacles.

Among the earliest trains to be described were the Rapakivi train,[100] the dolerite train from Corstorphine Hill, Edinburgh,[101] and some trains in the Rhône basin[102] and in Massachusetts.[103] These included the celebrated Richmond Boulder train[104] which crosses valleys and ridges diagonally as a fan of angular material plucked by the last ice. Other well-known examples are the peridotite boulder-fan of Rhode Island[105] and that from Iron Hill, Cumberland, U.S.A.[106] (fig. 66) where the outward movement from the centre line is about one-tenth of that directly forward. These and other trains in New England, illustrated by R. F. Flint,[107] are more numerous than in central North America where bedrock types are less diverse and the drift mantle is thicker. The Lennoxtown train[108] in central Scotland proceeds eastwards in accord with striae, roches moutonnées and drumlins from an essexite outcrop, 2,100 ft (600 m) by 300 ft (90 m), at the base of the Campsie Fells. Its axis is along the Firth of Forth and its northern boundary runs

by Carbert, Rosyth, North Queensferry and Burntisland, its southern one by Kilsyth, Bo'ness, Queensferry and Crammond. Fifteen miles (*c.* 24 km) from the source, the fan is 2½ miles (*c.* 4 km) across, in the longitude of Crammond, 40 miles (*c.* 65 km) across (Fig. 67).

The spreading and wearing out of the boulder trains cause the erratics to be zonally distributed in the drifts.[109]

Erratics and lines of flow. Lines of flow are ascertained easily if the relief is pronounced but much less readily on open plains. Here erratics are invaluable. Unless the rocks are uniform over wide areas, as among the Icelandic basalts, they give the direction better than, though generally speaking in entire harmony with the striae. These portray the basal flow at a particular point or a particular time only—usually the last glaciation (see p. 248)—while erratics give the mean or general direction, i.e. the resultant of all movements between the source and the site of the erratic.

FIG. 66.—Iron Hill boulder train, Rhode Island, U.S.A. W. H. Hobbs, 769 (2), p. 306, fig. 332.

As "indicator boulders" (Ger. *Leitgeschiebe*; Swed. *Ledblock*) igneous rocks, notably intrusive types, are generally the more suitable,[110] since the outcrops of sedimentary rocks (these have been preferred in Holland[111]) are usually larger and less well defined and have been eroded to an unknown extent along their margins.[112]

Intercrossing. The fact that erratics do not necessarily move on parallel lines but sometimes cross has often been commented upon since intercrossing was first observed in the track of the Rhône Glacier[113] and on the North German Plain.[114] On the submergence hypothesis, such mixing is self-evident. On the land-ice hypothesis it may be explained in innumerable ways (their relative and absolute importance is indeterminate and varied in space and time) which include derivations from (*a*) older conglomerates[115]; (*b*) pre-glacial stream gravels,[116] e.g. pebbles in the Thuringian Forest (gathered from "preglacial" terraces in the Unstrut), Elbe pebbles in Holland,[117] Bohemian pebbles in Mark Brandenburg,[118] Pliocene gravels with Silurian material transported from the Baltic region to Sadewitz in Lower Silesia,[119] or amber erratics in north Germany and England (between Yorkshire and Suffolk), obtained possibly from the *Alnarpsflod*[120] (see p. 281); (*c*) preglacial beaches, as postulated for Green Bay, Spitsbergen,[121] for flints in the Bornholm drift,[122] and for some pebbles in the

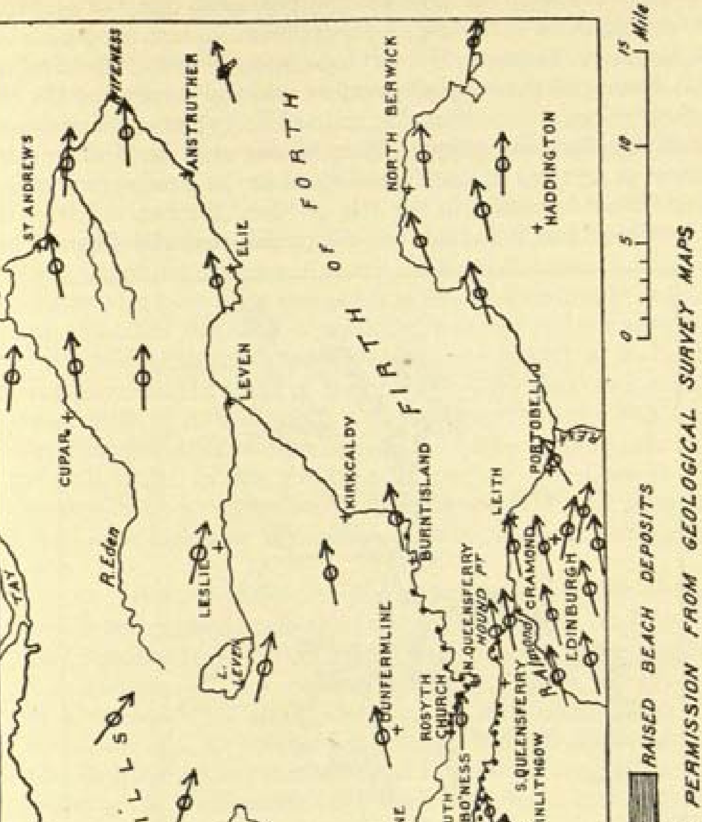

FIG. 67.—Lennoxtown boulder train, Midland Valley, Scotland. A. M. Peach, G. M. 1909, p. 29.

Holderness drift, including Scandinavian pebbles which had undergone marine erosion before glaciation[123]—the agreement in the percentages of boulders and pebbles, however, renders this doubtful[124]; and (*d*) subglacial streams in the case of small pieces. Others are connected with variations in the direction of flow due to (*a*) polar displacement[125]; (*b*) migrating

icesheds[126] (this changed the flow over Russia and transported Norwegian erratics to Britain at an early stage); (*c*) tectonic movements, such as those suggested for north Germany[127]; (*d*) ice-erosion[128] (doubtful and rare); (*e*) waxing and waning of the ice; (*f*) overflow of local centres by the main ice-sheet or their re-assertion after its retreat[129]; (*g*) one glacier overriding another parasitically (see p 87); (*h*) cross-currents at different levels in the ice[130] (see p. 272), as in the Vale of Eden or the overlapping trains of the Foxdale and Oatland granites in the Isle of Man, the two outcrops differing in height by about 400 ft (*c.* 120 m); (*i*) erratic material slipping on to the

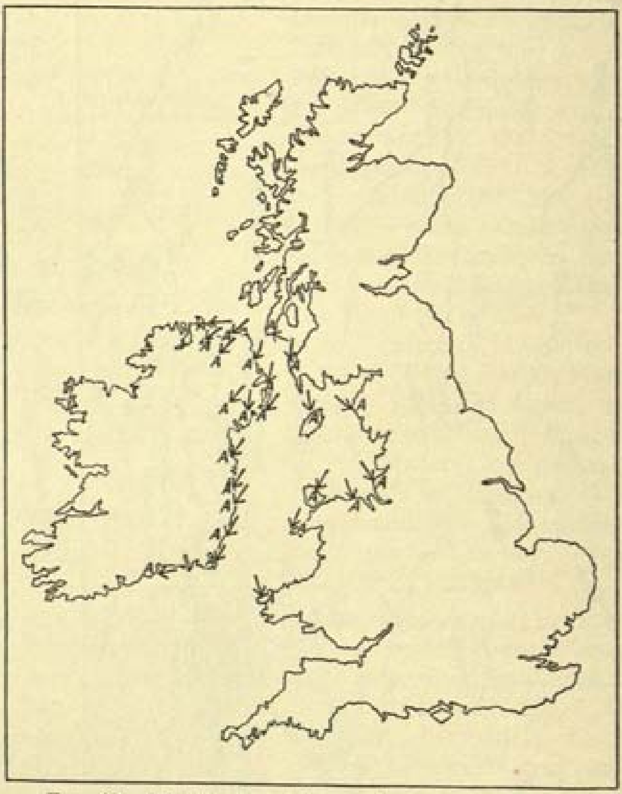

FIG. 68.—Distribution of the erratics from Ailsa Craig.

ice; and (*j*) moraines mixing in narrow passes which compressed the stream-lines.[131] Material may be incorporated from earlier drifts arising from these causes, as in the occasional Norwegian erratic of rhomb porphyry[132] in north Holland, east Friesland, Oldenburg, Denmark and Fyn, though later transverse movements sometimes scarcely disturbed the erratic fans.[133]

A fruitful cause of mixing was drift-ice, including bergs and ice-foot, which at the mercy of wind and tide scattered debris indiscriminately over the sea-floor during the early stages. This debris was incorporated in the ice which invaded the land from the sea, e.g. some of the Norwegian erratics[134] in the North Sea and in the drift of eastern England and of the Baltic; these include innumerable stones bored by *Saxicava* and *Clione* and well-rounded beach pebbles.[135]

The Ailsa Craig microgranite in the drift around the Irish Sea[136] (it was

found first in the Isle of Man[137] and subsequently in Dublin Bay[138] and all round the Irish Sea as far as Pembrokeshire and Tramore Bay,[139] at Silecroft in Cumberland[140] and in south Kintyre) may constitute a further example (fig. 68). Thus these pebbles are occasionally encrusted with millipores[141] and are associated with blocks of Carboniferous Limestone, coated with hardened shelly sand and scored by striae, and with erratics which could not have been transferred to their present sites by any known glaciers, e.g. serpentine at Birkenhead,[142] Shap granite and Borrowdale lavas in the Isle of Man,[143] and rocks resembling those of the Inner Hebrides[144] found in the Isle of Man, Cheshire, Anglesey, north Wales and north Cardiganshire.

Erratics may also have been derived from sources no longer exposed because the outcrop or outlier has been completely removed and the supply exhausted (an error to which thin sandstones are specially liable), as in the case of the Jotnian Sandstone in the Gulf of Bothnia,[145] of Magnesian Limestone found several miles north of its outcrop in Co. Durham[146] or of Chalk and its fossils in the Scandinavian drift of that county.[147] Alternatively, the outcrop was sealed beneath drifts in the progress of events—the Lias of Cheshire,[148] certain Scandinavian ore-deposits,[149] and diamonds[150] and "float copper"[151] south of the Great Lakes in Wisconsin, Illinois, Indiana, Ohio and Michigan are examples.

Outcrops on the inaccessible sea-floor were a more fertile source of error. This source is commonly neglected for historic reasons (icebergs could only receive their burdens from higher lands) and because we are ignorant of submarine distributions. Thus basalt, porphyry, upper Jurassic and Neocomian in Denmark,[152] Cambro-Silurian in north Germany,[153] the East Baltic Limestone and Leptaena Limestone,[154] Orthoceras Limestone and most of the Beyrichia Limestone[155] and Palaeocene and other sedimentary erratics in the north German drifts[156]—all these have been eroded from the Baltic in agreement with the flow-lines in Baltoscandia.[157] The "Brown Porphyry",[158] like the "Red Baltic Porphyry"[159] (= Rödo porphyry), was derived from south of Åland and some of the Rapakivi granite from the bottom of the Gulf of Bothnia.[160] Other erratics, including the Dalarne porphyry,[161] sustain this conclusion[162] and are gradually enabling us to construct a geological map of the Baltic sea-floor.[163]

The bed of the North Sea has in like manner supplied Neocomian and Gault erratics[164] to Schleswig-Holstein and Jutland; Chalk[165] to the drifts of Caithness, Newcastle and Scarborough (20 miles (32 km) north of the Chalk cliffs); flints[166] to the drift north of Wansbeck in Northumberland, at Newbiggin and Cleadon in Co. Durham, and to the Scandinavian drift at Warren House Gill; black flints, *Belemnitella lanceolata*, non-British cephalopoda and green-coated flints to the Holderness drift,[167] together with remanié fish remains from the Red Crag; Red Crag shells to Aberdeenshire[168]; and at Sheringham, a block of Newbourn Crag of a type unknown elsewhere.[169]

The Irish Sea has furnished many similar examples, the most important being Cretaceous flint.[170] This has been found in the surrounding drifts to St. Davids and Cardigan in South Wales and Ballycotton Bay, Co. Cork, and in the intermediate areas, e.g. at Ravenglass, over Cheshire and the English Midlands as far as Warwick and Bridgnorth and along the Irish coast from Co. Antrim southwards. It has occasionally been referred to east England[171] but usually and notably in the older literature to Antrim outcrops,[172] in seeming unison with the baked varieties and with chalk pebbles showing contacts

with olivine basalt.[173] A submarine source,[174] either Chalk or Tertiary deposits, is however suggested for some by their rusty gravel stain[175]; by their water-worn appearance and small size and the modes of percussion of beach pebbles[176]; by their increase in size and number as they are followed southwards from Dublin to Wexford[177]; by the lignites and coal fragments in the Wexford Gravels, derived from some submerged Tertiary and Carboniferous outcrops[178]; by the discovery, east of Pendine, in South Wales of sands with andalusite, tourmaline, cyanite and staurolite[179]; by the absence of Irish and Manx erratics in the Cheshire drifts—the vesicular basalt of Point Scarlet, Isle of Man, has, however, been identified in the Wem district[180]; and by the rocks and fossils[181] (from higher horizons than occur *in situ* in the Irish Lias) found in drift at Ballintoy and Ballycastle, Co. Antrim, in Dublin Bay, at Blackpool, near Chester, at Rochdale, Codsall, Wolverhampton and Strethill; and by the fossiliferous Middle Oolite erratic at New Brighton.[182]

Errors may be introduced by ballast[183] unloaded on shores and in rivers, e.g. French gravels with palaeolithic implements near Tönsberg, Norway,[184] Welsh and Danish rocks in Nova Scotia,[185] and Shap granite at Gainsborough in Lincolnshire.[186] Pebbles may also be derived from shipwrecks (laurvikite boulders near Dunbar[187]) or carried from the beach with manure on to the fields.[188] Little pebbles may be carried by seaweeds[189] or by sea-gulls, as in Maine.[190]

Uplift of erratics. The ice, which jettisoned most of its load at a low level, sometimes lifted the material higher, frequently much higher than the parent rock. Instances from the British Isles are numerous. They include blocks of greywacke near Kirkby Lonsdale,[191] Silurian rocks near Settle,[192] flints, Ailsa Craig microgranite and Eskdale granite on Moel Tryfaen,[193] and the 500 ft (*c.* 150 m) uplift of the Leavad erratic (see p. 363), the 600 ft (*c.* 180 m) uplift of Old Red Sandstone rocks in north Wales,[194] of certain rocks in the Lake District,[195] and of Permian Brockram and other rocks on the Pass of Stainmore[196]; and a 400 ft (120 m) uplift in the Wrekin area.[197] Similar observations have been made in Scotland,[198] e.g. in Galloway, Argyllshire, Pentland Hills, Central Highlands and Shetland Islands. Carboniferous gravels on the Wicklow Hills[199] and granite boulders on quartzite mountains in Co. Donegal[200] are Irish examples.

The uplift may amount to 1500 ft (*c.* 450 m), as in the case of the thrust Torridonian Sandstone west of Fannich Mountains,[201] and may be very rapid as instanced by the uplift of 1000 ft (*c.* 300 m) in 2½ miles (4 km) east of Loch Maree (see p. 668) and of 600 ft (*c.* 180 m) in 1½ miles (*c.* 2·5 km) south of Foxdale, Isle of Man.[202]

Uplifts are prone to occur if the iceshed was eccentric, as in some of the cases just mentioned from the Scottish Highlands and the 1000 ft (300 m) uplift of granite erratics west of the Rannoch Muir[203] and especially in Scandinavia (see p. 668). They were extensive south of the Baltic[204]—the Brown Baltic porphyry was lifted 800–1000 m—in north-west Russia[205] and to some degree in the Alps.[206] In North America,[207] where L. Vanuxem and J. Hall early noticed that the drift commonly occurred at elevations considerably higher than their source outcrops, Laurentian boulders were scattered plentifully south of the Laurentian region and over the plains and high land of west Canada, up to 5280 ft (1610 m).[208] Uplifts of 3000 ft (*c.* 900 m) have been recorded on Mount Katahdin[209] in Maine, of 5000 ft (*c.* 1525 m) on the

stoss side on Mount Washington,[210] and of 5300 ft (1615 m) in the Winnepeg area[211] and up to 4000 ft (1220 m) A.S.L. within 5 miles of the Rocky Mountains front.[212] Fragments of Potsdam Sandstone were carried from low levels near the north end of Lake Champlain to the tops of the Adirondacks.[213] Marine shells in the north German[214] and British drifts (see p. 630) are a special case.

Method of uplift. Much has been written of the way in which erratics and shells have been carried up. Many hypotheses are now only of historic interest; such are the waves of translation[215] (see p. 618), ice-foot,[216] anchor-ice,[217] seaweeds in a glacial sea,[218] solar radiation on stones encased in the ice,[219] and unequal postglacial upheaval[220] due to a rise of the isogeotherms[221]—this is inconsistent in Germany with the slope of the preglacial river terraces[222] and in Britain with the rate of erratic uplift (see above). Preglacial beaches may occasionally have provided material.[223]

Other discarded means are transport on an inclined plane from hypothetical great heights, implying enormous subsequent denudation,[224] and the overturning of bergs whereby basal debris was raised 300 m or flung even higher.[225] Darwin's hypothesis of flotation by pack-ice during progressive subsidence[226] involved an improbable longevity of the bergs or an extremely rapid subsidence. It was inconsistent with the diminishing size of the erratics as they are traced from their source and their discovery far above the limit of submergence postulated on other grounds.

Erratic uplift is possible, as is now generally conceded, by moving ice and is only part of the wider question of the uplift of glacial material, including marine shells. It is proved not only by the subglacial debris on the surface of glaciers, as frequently noticed during the past century[227] in Norway and the Alps and on the impact side of nunataks[228] in Spitsbergen and Greenland, but by the dyke-like ridges of debris squeezed up along transverse fissures[229] in Greenland and Spitsbergen, and by the inclusion of water-worn pebbles in shear-planes[230] and of marine shells in ice and moraines of arctic glaciers[231] up to 176 m A.S.L. The Sefström Glacier, Spitsbergen, advanced from its 1882 position (see K. Gripp's map[232]) and incorporated the shells of the sea-floor in its shelly moraine on Cora Island.[233] In this way, it affected an uplift of 30 m, possibly 75 m,[234] since a derivation from raised beaches[235] is inadmissible. The Von Post Glacier similarly incorporated marine shells in its moraines.[236] Marine serpulae and diatoms lie at 97·5 m on the slopes of Mount Erebus[237] and raised muds east of McMurdo Sound have been elevated through 100 fathoms (183 m).[238]

Effect of obstruction. Uplift by ice is analogous with that known to take place in rivers, especially in the base.[239] Obstructions retard the lower layers and compel a slight diagonally upward flow.[240] Debris also rises in crevasses and moulins,[241] by compressing waters in subglacial tunnels during the melting season,[242] and by upthrust at the snout.[243] Ice has been seen ascending gentle slopes (see p. 122). Moraines, ice-pedestals, masses of dead ice, frozen and rigid material or rock ridges projecting upwards into or as nunataks through the ice deflect the flow upwards,[244] provided the surface of the ice inclines more steeply than the reversed floor (see p. 123). This upward surge of the basal layers has been seen in modern glaciers,[245] attended by thrusting and gliding along shear-planes, and has been experimentally demonstrated on pitch and paraffin wax models.[246] Confirmative are the subglacial debris in

the contact planes of two glaciers (see p. 407), the impact sides of roches moutonnées, the ascending striae on rock-faces, the steep and irregular lee slopes,[247] the trains of local angular boulders in the lee of crags, and the transport and uplift of erratics across high ridges, as in the Shetland Islands.

Whether or not the ice and its burden descend again depends upon the relation between the vertical and flowage pressures[248] (cf. p. 253). This readily explains the contrasted form of the lee slopes of roches moutonnées and why some transverse ridges intercept distant erratics[249] while others allow so much material to cross that it distinctly colours the drift.[250]

The raising of material probably took place within definite but unknown limits; it was opposed by frictional melting at the base and by descending surface and englacial waters.[251]

F. Debenham[252] described a new method of ice-uplift as illustrated by the raised sea-muds in South Victoria Land (see above) which contain serpulae, foraminifera, polyzoa, corals, echinoid spines, sponge spicules, headless fish and deposits of mirabilite (sodium sulphate), the bulk of which was excessively fragile, well preserved, and often in the position of growth. He regarded this material as sea-bottom which was raised by the sea freezing at the base of the ice-sheet in contact with the sea-floor and the thawing of the upper surface of the ice. By repetition of the process, the marine muds ultimately emerged.

This method, it is objected,[253] is hardly compatible with the melting and depletion of the base of Ross Barrier that seems to be proved among other ways by the positive additions to the surface of that ice and by the firn composition of the Antarctic barrier bergs throughout. Nevertheless, Debenham applied the process, which operates in the Antarctic over considerable areas and with ice up to 213 m thick, to the Sefström Glacier (see above), to the British shelly drifts (see p. 630) and to the sea-ice off the polar coast of Ellesmere Land.[254]

Perched blocks. Surface erratics, such as those of the bouldery New York fields, are either the wreck of the drift whose finer materials have wasted away[255] ("erratics grow"), possibly by solifluxion, or they were deposited in this position as the ice retreated. They include the perched blocks in the Pyrenees[256] and in the Alps where de Saussure[257] noticed them. They do not denote, as was suggested,[258] that the valleys were filled to this level with detritus which was afterwards removed. They are poised precariously and apparently insecurely on the edges of cliffs or mountain sides and were let down gently at the dissolution to their present positions of precarious equilibrium. The "rocking stones" (Ger. *Wackelsteine*), which are so delicately balanced that they rock if pushed, are a type common in many places in west Ireland and New England. Many that were once perched on hillsides have doubtless crept or fallen to the valley bottom as in modern Greenland.[259]

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Chapter XVIII

BOULDER-CLAY

Boulders. Boulder-clay (Ger. *Geschiebelehm*; Fr. *argile à blocaux*; Dut. *keileem*) or till, a Scottish word[1] for "a kind of coarse, obdurate land", is physically and lithologically heterogeneous. Its boulders may vary from 1 to 99% of the accumulation and when examined statistically according to size-groups, based on the product of the three principal dimensions, may give a mathematical series.[2] They range from pebbles to boulders of great dimensions, according to the distance through which they have travelled and the hardness and structure of the mother rock—rocks which lend themselves to plucking provide a larger proportion of blocks while those which are readily abraded provide much clay, e.g. on Gotland. Chamberlin[3] suggested the following grading: pebbles, up to 15–20 cm long, boulderets, 20–38 cm, and boulders exceeding this length.

The boulders, using the word to embrace the several grades, often have a distinctive shape[4] though this mostly results, especially in the harder rock-types, from inheritance or the mode of derivation from the parent rock. If well glaciated, their plan may be roughly triangular or pentagonal with rounded and smoothed edges, humped back and smooth, flat sole. They then resemble a flat-iron or the prow of a flat-bottomed boat; the scour-snubbed nose is pointed by friction and the averted end is rounded, sub-angular or hackly. Friction along the bottom or, as Charpentier[5] noticed, along the sides of a glacier has faceted the boulders, some being ground perfectly flat. Such boulders (Jaekel's *Eiskanter*[6]) are flatter and less sharply angular than those of sand-blasted gravels[7] and have lateral facets running off towards the snubbed point. The larger the grip of the ice, the firmer and steadier was a boulder's course. A boulder with curved sole rocked a little as it advanced while rotation provided more than one facet, a number of them indicating alternate fixity and change of position, frequently repeated.

Although faceted boulders are common in erosive centres, as in Britain and Scandinavia, and occur in modern moraines[8] and Antarctic bergs,[9] they are relatively few near the periphery of the ice-sheets; only odd ones have been observed south of the Baltic.[10] This may account for the erroneous views concerning them which have been held in these countries, though E. Philippi[11] correctly diagnosed them; they have been referred to rotating pack-ice[12] and the repeated overriding of subglacial deposits[13] or regarded as wind-faceted pebbles[14] or fragments which were shaped before incorporation in the ice[15]—the drift, for example, in North America[16] does contain undoubted ventefacts which have been incorporated.

Striations like those on the bedrock are engraved on boulders at all depths in the clay. Especially liable to occur on limestones,[17] they may be confined to one or more facets or may cover the boulder, passing round its curves or crossing each other if it has shifted its position. On large boulders they are usually parallel with the longer axis as Darwin[18] observed.

Boulders, as Brongniart[19] noticed in 1828, are frequently horizontal or canted slightly downwards upstream, an adjustment beneath the ice recalling

the gentle impact slopes of roches moutonnées. The parallelism of boulders and nests of sand has been referred to "fossil" glide planes comparable to those of living glaciers[20] (see p. 117).

Boulders, except in narrow valleys athwart the ice-flow (see p. 323), also tend to be parallel[21] and in line with the direction of transport as given by striae on the underlying rock[22] (fig. 69): pebbles and boulders, for instance, prove the radial flow of the Oder Lobe.[23] Boulders on modern glaciers also tend to assume a longitudinal position.[24] Fabric patterns from successive layers of till sometimes record shifts of ice-flow similar to those shown by intersecting sets of striae on the bedrock[25] (see p. 901).

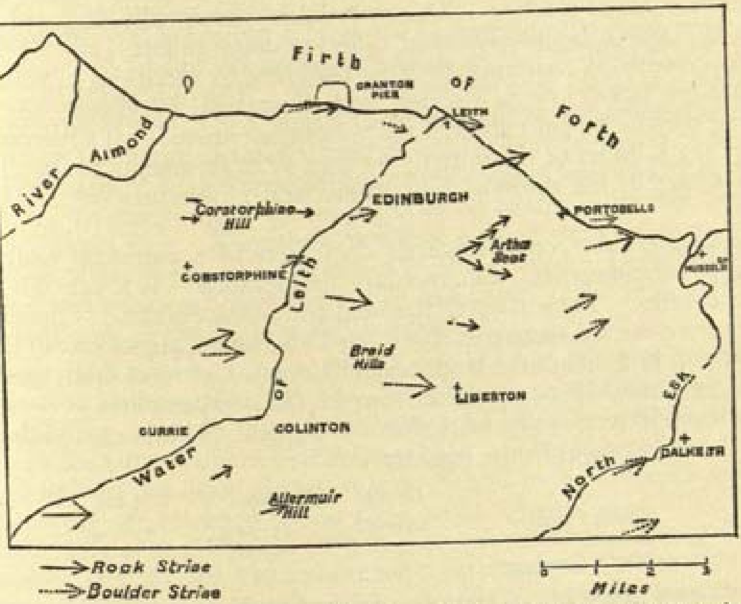

FIG. 69.—Parallelism of the major axes of boulders in the till with the striated surface on the rock *in situ* underneath near Edinburgh. H. Miller, *1904*, p. 165, fig. 1.

Matrix. Boulder-clay is typically tough, compact and tenacious, owing to the fine rock-flour it contains and the weight of the superincumbent ice. In the Oxford area and in the Thames basin it tends to be more resistant than the country rock, so that new valleys are prone to be eroded along its margins.[26] The texture varies with the source; on sandstones, as on the Potsdam Sandstone of Wisconsin or Britain's New Red Sandstone, it is liable to be loose and sandy; on granites, gneisses and quartzose schists, e.g. in parts of Finland and Co. Donegal, Ireland, it is stony, coarse and gravelly and often hardly distinguishable from decayed rock *in situ*. In these cases, the term "clay" is less appropriate than in areas of limestone, clay or shale, which as in south Sweden or on the Skiddaw Slates of the Lake District, provide the typical boulder-clay. Subsequent events have modified the compactness; it is less, for example, above the limit of the lateglacial submergence in Sweden than below it,[27] and elsewhere has been sometimes increased by induration from lime derived from a calcareous floor.[28]

The colour naturally varies with the material. Thus in Britain it is generally dull grey on schists, blue on such clays as the Oxford Clay, dark

blue or nearly black on Carboniferous clays and coals, and red on the Old or New Red Sandstone. Like the texture and composition, it changes in successive boulder-clays gleaned from different sources. The dominance of illite in Swedish glacial and postglacial clays supports the view that the bulk of these clays was derived from Tertiary and Mesozoic clays.[29] The distribution of the various types in Scania[30] reflects the lithology of the regional bedrock as do the Nebraskan and Kansan tills of the Mississippi and Missouri valleys.

Since ice is unable to sort its material and deposits mainly by liquefaction unrelated to the size of the ingredients, tills form a pell-mell agglomeration, unstratified and structureless and with boulders of various sizes, form and composition set at random. The material is occasionally disposed in wavy layers as between boulder-clays of different composition,[31] the waves being due to pressure, to resistance by large boulders, to variations in velocity and transporting power, or to packing and successive accretions. A pseudo-bedding may be brought about if the volume is reduced or if a rearrangement follows the solution of the underlying rock or of fragments in the drift such as the Chalk in the drifts of East Anglia and Holderness (pl. XIIA, facing p. 353).

The dispersal by ice of minerals distinctive of a particular source is an analogous phenomenon. Sedimentary petrography[32] is a valuable adjunct to field studies. Thus in north England and south Scotland,[33] fluorspar and barytes were carried eastwards along the Tyne valley, augite east of the Whin Sill outcrop in Northumberland, brookite north-eastwards from the Cheviot granite, and hornblende outwards from the Galloway granites.

The material tends to be worn down in transit. This is exemplified by the following percentages in the boulder-clay[34]:

	Gravel	*Sand*	*Clay*
South Sweden .	29·9–38	15–34·5	35·6–47
Denmark . .	4–8	57–80	15–35
North Germany .	3·9–4·3	55·8	39·6

Sands and gravels. Water often flushed the finer particles out of the drift during transport and so left a sandy till. Sands and gravels, intercalated with boulder-clay, form nests, pockets or impersistent seams, sometimes distorted or ripple-marked.[35] Though sparingly developed if the boulder-clay is thin and has little matrix, as in mountain glens, they are plentiful in valleys or on lowlands, as in central England and in north Germany where O. Torell[36] noticed them, or in the peripheral drift in which their bulk may equal or surpass the clays. The till is merely the end-member of the stratified drift series and grades into this; for all till was deposited from the ice in the presence of at least a thin film of water.

The materials are sorted according to weight and size; light coal fragments over the British coalfields, for example, are often segregated into definite layers. The grains, which form an unusually rich assemblage, are sharp, fresh and angular, especially where they were derived from crystalline or metamorphic rocks, quartz showing a conchoidal fracture and other minerals fresh cleavage faces.[37] Well-rounded shapes occasionally occur[38] as in sands remanié from desert (Trias) sandstone in Cheshire.

The enclosed clay balls or pebbles, boulders and spherical or ovoid masses, sometimes coated with sand or shell particles,[39] resulted when glacier-streams eroded boulder-clay.

Sands and gravels may be of local origin, as in north Germany,[40] or in the Boston area of Massachusetts where more than half are only 5–10 miles (8–16 km) from their source.[41] At other times the proportion of more distant constituents is higher than in the associated boulder-clay.[42] The difference is seemingly connected with the velocity of the streams which transported the gravels and with the wear and tear of these under the ice.

The pockets or lenses of sand may be merely "boulders" incorporated from an overridden outwash deposit or more generally may result from a local or temporary concentration of melt-water. Like the boulders themselves (see p. 376) they may have been tilted by the contemporaneous forward motion of the ice.[43]

Distribution. This ground-moraine of L. Agassiz—the term has at times been extended to include material deposited at the melting[44]—is neither uniformly thick nor universal. It spreads widely over plains and lowlands, e.g. the prairies of west Ohio, north Indiana, most of Illinois and of Iowa, and lodges on the floors of valleys, save in erosive centres, e.g. in Scotland[45] and the Cordillera of west Canada,[46] whence glaciers of later phases may have removed it. Its sheets slope up the hillsides on to the lower cols but leave the steeper flanks free. Here they pass into a stony drift or loose rubbly debris, ill-circumscribed and more or less locally derived. Detached patches occupy hollows between bosses and ridges and form "tails" in the lee of hill spurs or long spits in the angle at valley junctions.

Source. Both boulders and enclosing matrix (as much as 90% in south-east Wisconsin and west-central New York[47]) are predominantly gathered from underlying rocks or from those which lie slightly back on the drift course. In the Ottawa district, drift more than 5 miles (8 km) from the Pre-Cambrian upland has little of this material,[48] and in Nova Scotia drumlins which generally are confined to the slate country have found sufficient adhesive material up to 2–6 miles (*c.* 3–9·5 km) from the boundary with the quartzites.[49] On uplands, the colour and texture approach those of ordinary rock waste. The assemblage becomes more varied from the ice-centre towards the zone of maximum accumulation where the boulder-clay depends less upon the subjacent rock and is more or less of a general type. The "foreign" constituents which provide a sample of all the rocks upstream are more plentiful in thick than in thin deposits and the "carry over" from outcrop to outcrop is less sharp.

The various strata have generally contributed in inverse ratio to the distances over which the boulder-clay has moved, though the width and relief of the outcrop and the nature of the rock make modifications. Far-travelled debris is apt to be scarce because it tended to lodge or wear out in transit—the heavy basal load ensured a high rate of attrition—and ice spread from the centre and originated much material marginally.[50] Generally, the bulk, say 75–90% of the boulder-clay, has not journeyed more than 50 miles (80 km). Long-distance erratics (see p. 124) may have travelled superglacially or englacially.[51]

When the ice came in from the sea, e.g. on the east and west coasts of England, it dredged up the shelly, submarine deposits, including the early-glacial droppings of bergs or drift-ice, and incorporated them in the drift of the land. Boulder-clay also contains rock-flour ground by the ice, a source which E. Collomb[52] first appreciated. Analysis shows that the rock-flour in

boulder-clay is essentially a quartz-flour[53]; this agrees with A. Daubrée's observation that the milky turbidity of the Rhine, even some hundreds of miles from the Alpine glaciers, is principally due to this mineral in impalpable form.

The "local moraine"[54] of Dollfus-Ausset and German geologists (Torell's "local ground moraine"[55]) is a rubble of local fragments[56] ("*in situ* moraine" of F. Svenonius[57]). Sometimes cemented into breccia where composed of limestone, it occurs at the base of the drift, often in sheltered localities, such as in transverse valleys or at the foot of steep cliffs, prominent ridges or ledges. Its glacial age is proved by occasional erratics or the disturbed underlying rock. It may be earlier than the overlying boulder-clay as Torell surmised or may have been deposited simultaneously with it,[58] by plucking at the sole of the ice or by frost shattering.[59] It may in places be an early glacial "head".[60]

Subglacial origin. Although the boulder-clay has been the subject of a vast literature since Agassiz[61] first described it in 1837, it remains to some extent the "mysterious deposit" R. Chambers designated it. It is an accumulation *sui generis*.

FIG. 70.—Fluxion structure in till (natural size) Fillyside, near Edinburgh. The arrow denotes the direction of ice-movement. H. Miller, *1904*, p. 178, fig. 8.

It was widely believed in the infancy of the glacial theory, though strenuously denied by the "submergers", that boulder-clay accumulated under the ice and was dragged along by it, being fluid enough to allow of motion.[62] The ice communicated its own motion to the clay, moving the bottommost layers very slightly, the upper ones at an ever increasing rate, yet extremely slowly towards the junction of ice and clay. By particle moving over particle, the boulders within the mass became intimately glaciated. Supporting this view of viscous drag were the "tails" in the lee and the "pre-crags" in front of projections; the boulder-clay's compactness, jointing and dominantly local character[63]; its occasional plane resembling cleavage slip[64] ascribed to pressure by the superincumbent ice (ice 1000 ft thick exerted a pressure of *c.* 30 tons/sq. ft); and its "burst boulders",[65] completely disintegrated or only partially broken, e.g. in the Dutch, Danish and north German clays, or crushed and rolled out flat or splintered into fragments and dispersed over a number of square metres. Some of these boulders are composed of material which readily absorbs water and have moved little if at all after bursting, the parts being in contact; they may perhaps be attributed to frost,[66] as on recent Spitsbergen moraines.[67]

Fluxion structure[68] may corroborate the subglacial view. This rude foliation or fissility, the "streaks" of C. Reid,[69] which characterises tills containing much fine material (fig. 70), is due to shearing or drag in the mass; the faces of the soled boulders are horizontal, the grains are orientated and striated, the material has flowed around big, unyielding boulders, and fractured boulders have their upper parts forward. Where rocks projected into the sole they caused schuppen structure, folding and wedging of the drift.[70]

The "striated boulder pavement" is cited in further confirmation. In a

mass of till the boulders on a particular horizon sometimes have parallel striae engraved on their upper faces. Though the smaller boulders may have escaped striation, the clay itself is occasionally striated.[71] These features, first described by C. Maclaren[72] in 1828 and subsequently by D. M. Home[73] in Scotland, by K. Schimper in Switzerland,[74] and by O. N. Stoddard in North America[75] (Miller[76] has given other early references), were afterwards noticed in many countries.[77]

Advocates of half-floated glaciers regarded them as made by intermittent pressure on accumulating material[78] while others ascribed them to fits and starts of subsidence,[79] to stages in a slowly growing boulder-clay,[80] to deposition at the base of the ice following pressure melting, or to subsequent glaciation by thin ice.[81] They more probably represent differential motion along shear-planes accompanied by horizontal slickenside.[82] They may mark an advance following temporary stagnation[83] or an interval of till erosion[84] (though not necessarily an interglacial one as suggested[85]); for the colour and composition differ occasionally in the clays above and below,[86] fossiliferous layers occur on the horizons,[87] and pavements are known in front of existing glaciers,[88] e.g. the Upper Grindelwald and Argentière glaciers of the Alps and the Illecillewaet Glacier in the Canadian Rocky Mountains.

Ground-moraine of present ice. Basal dirt or ground-moraine has been repeatedly observed under modern glaciers.[89] Hugi[90] was the first to note it; Agassiz termed it *boue glaciare*. It has been recorded from the Alps,[91] Sierra Nevadas of North America,[92] Greenland[93] and Spitsbergen[94] where typical boulder-clay is often associated with present-day ice. Himalayan glaciers have an unusually great amount of subglacial material—they often "swim" on it.[95] Yet basal dirt is frequently absent,[96] as beneath the glaciers of Mont Blanc, the Antarctic and cirque glaciers generally, or is very rare,[97] e.g. in the Alps and Greenland. If present, it is invariably thin (it is not more than 2 m thick in Greenland[98]) and is not true ground-moraine: S. Finsterwalder[99] has stated this for the Alps and Gripp[100] for Spitsbergen, since much is probably the lowest englacial layer of sand and mud with an ice-cement.[101]

This rarity of modern ground-moraine has been attributed to the removal long ago of the disintegrated material[102]; to the nature of the country rocks, e.g. in Greenland[103] which could only supply coarse and arenaceous debris; or to the impossibility of examining more than the margins of the ice: the uprise of material at nunataks, as in Greenland, proves that it exists in the inner parts.[104]

While basal debris is discoverable in some glaciers, it cannot be gainsaid that nowhere in accessible positions is *moraine profonde* forming to-day comparable in thickness with that of the Ice Age. It is even doubtful whether thick boulder-clays were transported subglacially[105] in Pleistocene time. While thin layers, facilitated by saturating the muds and sands with water,[106] were moved as evinced by debris extruded at the end of existing glaciers and the folding of subglacial rocks in glaciated areas beneath boulder-clay[107] (folding and deposition may have been successive), movement of thick masses is unlikely because the internal friction in the ground-moraine was great and the lessened plasticity in the lower layers when mixed with sand and clay compelled the upper ice to ride forward over a stagnant layer,[108] as is occasionally seen in modern glaciers.[109] The depth of till which can be dragged along subglacially is probably definitely limited[110] (N. S. Shaler[111] made it

30 m), though it may vary considerably according to the depth of the ice and the density, nature and degree of saturation of the debris. Glaciated rock under boulder-clay may indicate that erosion preceded deposition and not that the clay was dragged along simultaneously.

Englacial material. These considerations suggest that some till was carried englacially in a kind of ice-conglomerate or breccia in the glacier's lower strata and was thawed out during melting from the very lowest debris while that above, still frozen, continued to move on. The overriding layers imparted the toughness to the clay as it came to rest.[112]

Such englacial material, to use Chamberlin's term[113] ("inner glacial", "in-glacial" and "intraglacial" are alternatives[114]), though sometimes absent,[115] is, unlike the subglacial material, very common in all modern glacier-regions[116] though it is masked in the Alps by the abundant superglacial detritus which led Agassiz[117] to deny its existence except in crevasses. It is either uniformly distributed or more generally, as in Greenland,[118] collected into layers, lines or lenticles, parallel with the sole.[119] The lenticles dovetail into each other, the stratification being frequently extremely delicate and developing into lamination with as many as 8 laminae to 1 cm. The layers, plane or slightly undulatory, curve round the boulders with marked faulting, drag and thrust phenomena, especially towards the margin. Thrusts may traverse all the layers or carry clean ice over them.

The dirt in Greenland is confined to the lower 50, 75 or exceptionally 150 ft (15, 23 or 46 m)[120]. It is more abundant in ice descending cataracts or ice-cliffs[121] and, according to Chamberlin,[122] at the sides, according to Drygalski,[123] (who argued from the distribution in overturned bergs), in the middle of glaciers. Seismic methods[124] suggest it may be 15–30 m thick at a distance of 120 km from the edge.

There is often a complete transition from clear ice through ice with much englacial material and basal dirt (by some[125] included in, by others[126] excluded from the term englacial) to the underlying boulder-clay.[127] No sharp line can be drawn between englacial and subglacial debris though a distinction between basal dirt and ground-moraine is often recognised.[128]

Englacial material suffers little in transport.[129] Thus Forbes[130] early recorded the emergence, almost undamaged, of a bottle and knapsack which had travelled 1310 m in 10 years, while a geological hammer was recovered after 15 years, the iron without rust and the wood undecayed. Carcases of sheep reappeared intact[131] and a gauze veil melted out untorn from the Glacier des Bossons after 41 years in the ice.[132] Similarly, rock-fragments remain angular. Nevertheless, they are sometimes striated and rounded by mutual attrition,[133] mainly by shear along the debris layers, occasionally along the sole of a parasitic glacier,[134] englacial transport being differential and increasing upwards.

Englacial and subglacial detritus is derived not only from the margins and surface, as early emphasised,[135] but from other sources, such as inner moraines[136] and subglacial erosion.[137] This is proved by witnessing the actual operation[138] and in indirect ways. Thus englacial debris is found in glaciers which have no superglacial debris[139]; rocks which never projected above the ice occur in the drift, e.g. remanié fossils and detritus in the drift of eastern England which differ from those of the land outcrops,[140] Silurian limestones of Gotland, Chalk of the Baltic region[141] and the Richmond Boulder-train[142]; and marine shells are embedded in moraines in some

modern glaciers of Spitsbergen and Norway (see p. 631) and in boulder-clays in east and west England (see p. 630). Even "local moraines" (see above) may mean much glacial erosion.[143]

While, therefore, it is untrue to say that boulders could not be incorporated in the sole or if incorporated would be detached at the first favourable accident,[144] the precise mode of incorporation is less certain since ice is incapable of lifting debris through turbulence. The method of incorporation has indeed been much debated since Charpentier[145] attributed it to expansion on freezing and Forbes[146] invoked friction and frontal resistance. Chamberlin[147] was the first to point out the mechanism. Drift was transformed from subglacial to englacial level at the summits of subglacial steps or of projections into the ice where shear-planes, marked by bands of debris, developed. This, however, operated only if the ice was thin since thick ice brought the englacial material down again in the lee of obstacles to become subglacial once more.[148] Incorporation may also have been accomplished if a glacier overrode earlier glacial deposits[149] or frontal talus which fell from overhanging cornices,[150] the extraglacial detritus being wedged into the base[151] in the form of "banding" by pressure or glide planes or differential flow.

Preglacially weathered rock and soil, saturated with water, was frozen during the early glacial stages to an appreciable depth beneath the ice, possibly to its junction with the live rock, which was the then true base of the ice-sheet. Shear-planes developed in the unfrozen soil, if any existed, so that much debris in this way became englacial from the start.[152]

While we may be sure that the Pleistocene ice had such englacial debris, its thickness, which varied not a little,[153] is not known. It was probably far less than Dana's figure of 150 m[154] and of the order of that in modern Greenland where its depth remains constant as the ice-sheet thickens backwards from the edge[155] and is not less in the much smaller and thinner island-ice off the coast.[156] Moreover, no known process of absorption is competent to diffuse detritus through any considerable depth,[157] and pronounced irregularities of relief favouring upthrust of material were generally missing in the path of the ice-sheets of Europe and North America.[158]

Englacial origin. The englacial origin of boulder-clay was first sponsored by J. G. Goodchild[159] in Britain and about the same time by several American geologists.[160] Glacial studies in Spitsbergen and Greenland during the last decade of the 19th century (see above) gave it an impetus and led to its adoption in both the Old and the New World.[161] Some, including O. Torell[162] and other monoglacialists (see below), distinguished a lower boulder-clay, dense, tough and compact of subglacial origin (the "till" of Scotland and *pinnmo* of Finland), which alone deserves C. Martin's term[163] *moraine profonde* or ground-moraine,[164] and an overlying boulder-clay, separated from it by a distinct plane and sprung either from englacial (or superglacial) sources and let down as a superficial sheet at the melting,[165] or consisting of bottom dirt brought up higher into the ice by shearing movements.[166] A layer of stratified sand, formed by melt-waters, sometimes marked the plane of separation.[167] This differentiation may be justified if the upper clay is very thin and without an outer limit distinct from the lower clay but is unwarranted where the clays differ in composition. How much drift is englacial and how much subglacial is, however, problematical. Discriminating criteria are somewhat doubtful, partly because of postglacial alteration, including hydration and oxidation.[168] A superglacial origin for the upper till

is to be excluded where petrographic and mineralogical analyses show that it has the same stage of grinding or a different orientation of its grains. (B. Järnefors, 1952.)

N. O. Holst[169] and W. Upham,[170] both monoglacialists, among others fully discussed the question. They thought the englacial drift was 40–50 ft (12–15 m) thick and equalled the stratified drift, and contended that the englacial boulder-clay, the "upper till" of Torell, Hitchcock, Upham and Chamberlin, was coarser and looser in texture and more gravelly and sandy, a difference which led some, e.g. F. Johnstrup, to see in the upper till a product of drift-ice. The upper till contains "gutta percha" clays and thin beds of sand and boulders, the latter of more distant origin and bigger and less glaciated, more plentiful, and not definitely orientated. Yet loosely packed boulder-clay may be subglacial[171] and subglacial clay may be stratified if much early glacial debris was incorporated during the advance. In recent years R. G. Carruthers[172] has repeatedly emphasised the englacial origin of till and deposition by bottom melt—top melt has left little or no trace since the upper ice was dirt-free. The Firth of Forth, Tyne, Tees, Humber and Wash glaciers in his opinion overrode the North Sea ice as vast parasitic glaciers and were compelled by the pressure of that ice to ride forward on flat-lying shear planes.

Surface boulders such as granite, early attributed to an "Erratic period",[173] are frequently not only more angular[174] and larger, as on the plains of west Canada,[175] but were gathered from more distant sources than those of the subglacial drift.[176] They probably came from superglacial or higher englacial horizons,[177] sometimes from projecting nunataks,[178] as in parts of Västerbotten and Östergotland, though exceptionally bergs may have floated them as in parts of Finland.[179] Their considerable increase after a readvance may denote interglacial erosion of a till sheet,[180] particularly in interglacial valleys, or an interval of frost action.[181]

The arguments urged against an englacial origin,[182] namely, that the drift is local and the englacial material is thin and was enclosed only with difficulty, are without force. The load was probably much bigger than in modern ice-sheets,[183] though its vertical range, as just mentioned, was probably of the same order.

The interbedded sands and gravels may have been obtained from pre-existing alluvium, as in the Glasgow district,[184] from subglacial streams,[185] or by riding over valley trains and outwash[186] or book-leaf clays and other deposits of extraglacial lakes.[187] The many oscillations of the advancing ice-edge caused their aggregate volume to be considerable.

Gripp[188] concluded from his Spitsbergen observations that boulder-clay was melted out of ice penetrated by countless dirt-filled cracks. The dirt formed cones at the surface, as H. Backlund in Spitsbergen and Koch and Wegener in Greenland had already noticed.

Time and place of deposition. Boulder-clay may have been laid down during maximum glaciation in more sheltered localities, e.g. in transverse valleys (see p. 323), or in expansions of valleys which ran parallel with the ice-flow.[189] As its local stratification implies, it may have accumulated by adding layer upon layer. The bulk, however, was deposited submarginally,[190] say, 50 miles (80 km) within the ice-edge[191] (some would say marginally[192]), either under the ice[193] or englacially as this passed into the ablation zone and sheet after sheet of englacial material ceased to move: the contortions some-

times seen in the ground-moraine have been interpreted as structures preserved from the original ice[194] (cf. p. 398). The erosion of *Zungenbecken* (see ch. XII) and the deposition of boulder-clay including drumlins (see ch. XIX) were successive processes.[195]

Striated boulder pavements and the parallelism of larger boulders with the striae on underlying rocks (see p. 377) show that the ice was still moving. Distant erratics in the clays are no real obstacle to this view since some were unquestionably carried by subglacial streams or by floating ice during early stages either in the sea or in glacier-lakes[196] as in north Germany and certain parts of the British Isles.

The planes, which are either latent or rendered visible in the boulder-clay by lenses or bands of sand, etc., and which are horizontal unless subglacial irregularities caused the formation of schuppen structure, have been interpreted as shear-planes corresponding to the original structure of the ice.[197]

Till plains. Boulder-clay shrouds the country as broad, uniformly flat and expressionless plains over wide areas, either undissected or cut by shallow stream courses displaying a dendritic pattern. These *Grundmoränenebene*[198] are essentially smooth or gently rolling but have no definite alignment of undulations. They are found in north Germany,[199] e.g. north of the Baltic Ridge, in Posen and about Leipzig, over much of Finland,[200] in Holderness and East Anglia, Durham and Northumberland and central Ireland in the British Isles,[201] and in North America[202] in Wisconsin and the area about the Mississippi, e.g. in Illinois and Missouri, where the movement of the ice was controlled and directed by deep valleys.[203] Unevenness was introduced by epiglacial denudation, by subdued irregularities in the rock-floor, or by unequal deposition due to varying rates of supply. While the topography is apt to be guided by the subjacent rocks, if the boulder-clay is thin, thick drift is likely to induce its own topography,[204] as in north Sweden and east of the Baltic. It may give a mere striped appearance to the topographic maps or rise into irregular swells that pay no respect to the laws of symmetry.[205] Such plains (see p. 399) may owe their existence to the thickness of the drift,[206] to deposition by dead ice[207], to the slow and uniform rate of retreat, or to the later periglacial effects (see p. 1150).

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Chapter XIX

DRUMLINS

Form. Boulder-clay, which as a rule forms plains (see p. 385), gives rise sometimes to the peculiar "drumlin landscape".[1] Typical drumlins are among Nature's most graceful hills. Their smooth, oval mounds or elongated ridges possess straight major axes and have rounded summits, steep flanks, regular contours, and in profile flattened double sigmoid curves. The shape was faithfully reflected both in the earliest scientific terminology,[2] the "parallel" or "elongated ridges" and "mammillary" or "elliptical hills", and in such popular descriptive words as "hogbacks" and "horsebacks" in the interior and "whalebacks" in the coastal regions of the U.S.A.[3] The term drumlin, from *druim* a mound or rounded hill, a gaelic word extremely common in place-names[4] in Scotland and Ireland, was used by J. Bryce[5] in 1833 and brought into glacial literature by Close[6] in 1866: W. M. Davis[7] introduced it into America.

Drumlins occur usually *en échelon*, sometimes serially in ranks side by side and very occasionally as solitary features. They may merge at their base or dispose themselves at moderate intervals. Exceptionally, they are double-tailed or double- or triple-crested, the subordinate crests overlapping *en échelon*[8]; drumlins with undulating crest-lines or a slight sagging in the summit are intermediate stages.[9] Such twin-, triple-crested or serrated drumlins grew from mounds which were initially separate. In some districts, the drumlins are grouped together in vast numbers, their tops fusing into a fairly level drumlin upland.[10]

Alden[11] has well described how greatly the drumlins vary in their breadth, length and height. Keilhack[12] recognised five and T. C. Chamberlin[13] the following four types: (*a*) lenticular or elliptical hills; (*b*) elongated ridges; (*c*) mammillary hills; and (*d*) till tumuli or immature drumlin nuclei ("pseudo-drumlins"[14]). They grade from conspicuous hills, 200 ft (60 m) high, as in New York State,[15] to a subdued drumlin topography or "drumlinoid surface",[16] or to faint, scarcely noticeable swells. These low, broad mouldings or indefinite flutings form Fairchild's "washboard structure"[17] which only a striped appearance of the contours or parallelism of the minor streams suggests.

Compilations[18] of published statistics show that the average drumlin is less than 1000 m long and has a length-breadth ratio of about 2·5:1; in north Germany this ratio is 3·75:1.[19] The mean height is not so commonly recorded: it is less easily judged in the field or read from maps. Generally speaking, it is relatively small, rarely exceeding 20 or 50 m[20]; the vast majority of German drumlins are 5–15 m high[21] though the average in New York State is 180 ft (55 m).[22] According to Früh,[23] the size and mass are inversely proportional to the number and distance apart of the individuals.

Notwithstanding their diversity, drumlins are governed by certain laws which are generally though not absolutely valid. In one and the same region, as in west and north-east Ireland, Upper Bavaria, British Columbia, Nova Scotia and New York, they approximate (much more so than do roches

moutonnées) to a uniform size[24]; subtypes are not intermingled and a particular type occurs almost to the exclusion of any other. Thus in New York, the mounds are usually elongate and much longer than in Wisconsin[25] where in the west they are higher and narrower, more crowded, and more steeply sloped than in most of the state,[26] though in the east they are prevailingly circular and dome-shaped—they are the original mammillary hills of Chamberlin. They are practically circular in parts of Massachusetts but are oval near Boston[27] (fig. 71). In Maine,[28] they are lenticular in the west and elongate in the east and smaller than in New York, New Hampshire or Massachusetts. Those in north Saskatchewan and Manitoba are remarkably long and narrow

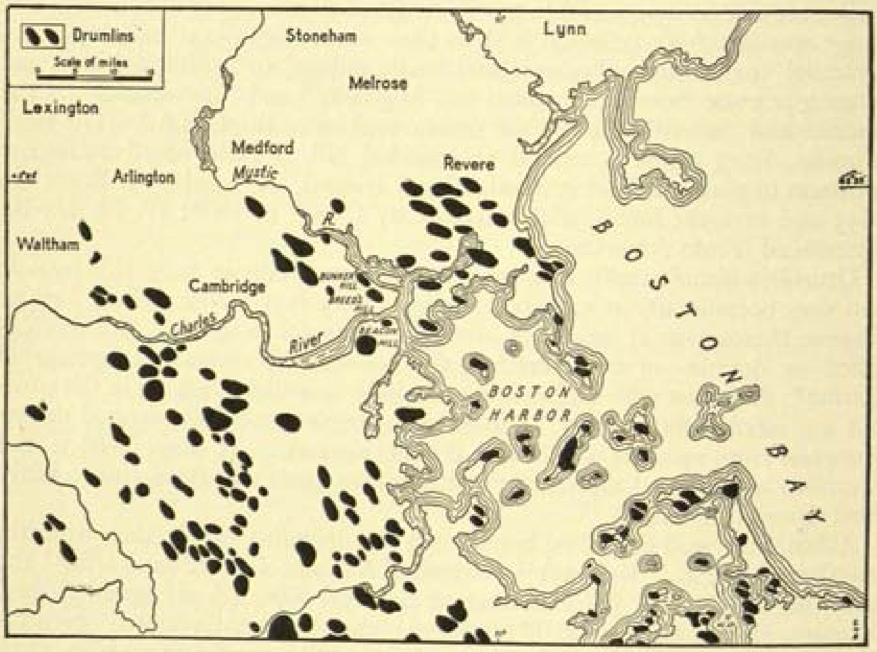

FIG. 71.—The larger drumlins in the vicinity of Boston, U.S.A. W. W. Atwood, *83*, p. 90, fig. 43.

and made of very sandy till: they were early referred to by the Indian name *ispatinow*.[29] Finnish drumlins belong in general to Chamberlin's type of elongated ridges[30] (pl. XIIIA, facing p. 416).

In any one field, the drumlins change from the margin inwards and in the direction of flow. Thus they lengthen notably towards the axis of the lobe in Wisconsin[31] and in New York State[32] where they are rounder in the north and long, low and narrow in the south; in the Finger Lakes region they are longer in the north and smaller and lower in the south.[33] In Upper Bavaria,[34] they are narrower and partly longer near the margins, and at the edge of the drumlin field of the Inn valley become imperfect and embryonic, i.e. they make slight waves with drumlin-like lines and curves.[35] Traced into the uplands, drumlins tend to become bigger, less regular and fewer; they merge into gentle undulations and flutings[36]—the north Swedish *Lide* are of this type.[37] At the margins, they pass imperceptibly into rock-drumlins (see

below) or scatter out upon the plain[38]; they also fade into faint swells or flutings of the till, one or two metres in height.[39]

Composition. Drumlins generally consist of the local boulder-clay. They sometimes enclose stratified material,[40] notably at the base and in the core,[41] or if they rest on outwash[42] or on the granite of the Canadian Shield[43]; they may indeed consist of nothing else,[44] as in the Ill valley of Austria or about Kelso in Scotland. They sometimes display a quasi-stratification, often arched to conform with the curved surface,[45] and occasionally a proximal pressure structure and a distal "tip-bedding".[46] Bedding, however, is generally inconspicuous but is brought out by lines of boulders, by streaks of sand, or by seams of loam and stoneless clay. It appears in differences of texture, of weathering, and of colour connected with the various layers' capacity for moisture.[47]

The rock below drumlins is usually glaciated.[48] It crops out in the intervening hollows or is concealed under thin drift. It often forms a core,[49] though whether this is the rule or the exception it is impossible to say. Cores occur in Fennoscandia[50] (they are typical in north Sweden[51] and in the larger mounds of Närke[52]), in Jasmund and Rügen,[53] and in Switzerland,[54] but are rare in Latvia,[55] in the Erie and Ohio basins,[56] in east Wisconsin,[57] and in New Hampshire and Massachusetts.[58] About Boston, the drumlins appear to be aligned above bedrock ledges.[59]

The core's position relative to the contour of the hill varies very much.[60] The coat of drift may rest uniformly and symmetrically on the rock-nucleus as in the "veneered hills"[61] or be eccentric to it, the nucleus in this case being at the crest or, in most instances, towards the impact or proximal end. This is noticeably so if the impact faces are very steep, as in north Sweden[62] where drumlins with cores completely covered pass into forms which had only enough drift to add "tails" to the roches moutonnées.

The core may swell until the whole drumlin is wrought in solid rock as Close[63] observed. These "rock-drumlins",[64] "rock-drums", "drumlinoids"[65] or "false drumlins"[66] (Ger. *Rundhöckerdrumlins*[67]) are sometimes found in the proximal end of drumlin fields.[68] Their shape is almost indistinguishable from true drumlins if they coincide with the strike of the rock[69] but is generally less symmetrical or regularly sloped, is larger and narrower,[70] and has an impact face which is liable to be more abrupt and uneven.[71] Transitions proving a common origin link rock-drumlins with true drift drumlins on the one hand[72] and on the other with roches moutonnées[73] and crags and tails.[74]

Distribution. The conditions necessary for drumlin building were apparently seldom realised: drumlins occupy only a small proportion of the till country. They are rare over vast regions of glaciated Europe, and in North America are infrequent in Pennsylvania and New Jersey and lacking in Ohio, Indiana, Illinois, Minnesota, the Dakotas, and south Michigan. They are distributed, apparently capriciously, over wide plains and valley expansions and at debouchures, and less commonly on slopes or uplands.[75] Packed in close-set series, they resemble a green, billowy sea, as in parts of Canada[76] or in the "basket-of-eggs" country of Co. Down, Ireland.[77] Partially submerged tracts, e.g. Boston Harbour, Massachusetts, have island-studded shores—many of the *inis* of Ireland, e.g. in Clew Bay, Donegal Bay and in Lough Erne, are drumlins (fig. 72; pl. XIIB, facing p. 353).

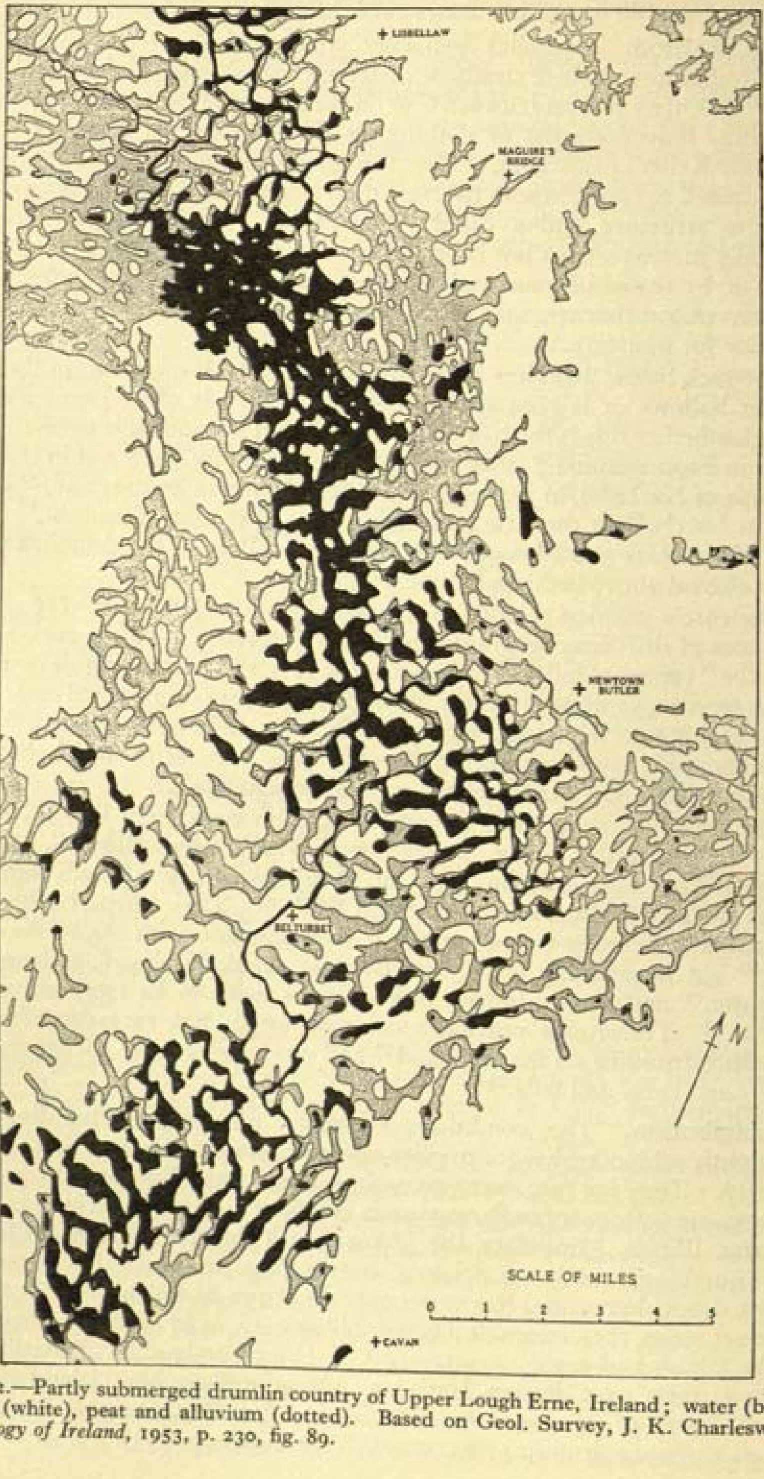

FIG. 72.—Partly submerged drumlin country of Upper Lough Erne, Ireland; water (black), land (white), peat and alluvium (dotted). Based on Geol. Survey, J. K. Charlesworth, *Geology of Ireland*, 1953, p. 230, fig. 89.

Drumlins sometimes dispose themselves in belts of considerable width which traverse flat plains of till athwart the ice-flow; examples occur in New Hampshire, Central New York and Massachusetts.[78] In North America, they often lie behind morainic lines[79] and in Switzerland[80] are prone to rise between morainic girdles and central basins or from higher ground between *Zweigbecken*.

British drumlins are especially well developed in Co. Donegal,[81] Co. Mayo[82] and in north-east Ireland[83] (see fig. 252, p. 1212) where they extend from the Ards Peninsula to the Shannon and Co. Louth and constitute one of the biggest continuous drumlin countries in the world; in Galloway,[84] the Tweed valley[85] and the Midland Valley[86] in Scotland; in Wensleydale,[87] the neighbourhood of Kendal, Oxenholme and the Ribble Valley,[88] between Hellifield and Skipton,[89] in the Vale of Eden and Solway area[90] in England; and in Anglesey and the Wrexham district in north Wales.[91] On the mainland of Europe, where they were long ignored (they were first described from the Salzach district[92]), they are found in various parts within the Alpine glaciation[93] and, relatively rarely, in north Germany[94] (Ger. *Rückenberge*), in Holland,[95] and in Jutland and Denmark.[96] They have been recorded from the lands east and south-east of the Baltic[97] and from Poland,[98] the Ukraine[99] and Fennoscandia[100] (e.g. Närke, Östergotland, Västergotland, Finland). Forms have been identified with them from central France,[101] the Dinaric Alps,[102] Tienshan,[103] Siberia,[104] Novaya Zemlya[105] and China.[106]

North American drumlins are grouped mainly in five areas: (*a*) Manitoba and Athabasca[107]; (*b*) New England, ranging from Ontario,[108] New Brunswick[109] and Nova Scotia[110] (2300 in the south-west) through south Maine[111] and New Hampshire[112] (*c*. 700) to Connecticut and Massachusetts[113] (*c*. 1800), including 180 about Boston; (*c*) Michigan and Wisconsin,[114] of whose 5000 drumlins 1400 are situated in the south-east; and (*d*) central Western New York.[115] This state has one of the most remarkable groups in the world; the belt which is 35 miles (56 km) broad and 140 miles (*c*. 225 km) long between Lake Ontario and the Finger Lakes (fig. 56, p. 284) and rises to 1700 ft (518 m), comprises 5000 sq. miles (*c*. 13,000 sq. km) and over 10,000 drumlins. Where set close, they are 20–35 per 4 square miles, though 5 to the square mile is common and 3 is the average (cf. Baden, where they are 6 per square kilometre[116]). A fifth area occurs in British Columbia[117] where they probably number several hundred thousand.

Original features. The origin of drumlins has long been a vexed question; many hypotheses, such as those which associate them with the sites of ancient moulins[118] or rock-streams,[119] require no refutation to-day. Equally indefensible is the assertion that drumlins are not moulds impressed by ice but remnants of uniform and once-continuous sheets dissected by tidal or marine currents,[120] by rain and streams,[121] by extraglacial waters at the retreat[122] or by a reticulated subglacial drainage.[123] But postglacial modelling is inconsistent with the arched bedding, the parallelism with the ice-flow, the transition into roches moutonnées or crags and tails, the lack of boulder-strewn tops and the absence of drumlins from much of the boulder-clay country or from Mesozoic or Tertiary clays.[124] Drumlins preceded the deposition of moraines,[125] loess,[126] kames, kame-terraces and osar,[127] the osar sometimes side-stepping in a zig-zag course across them[128] or extending each way from a traversing trench[129] or running in the troughs between the drumlins and parallel with their major axes.[130] They antedated lateglacial lakes,[131]

as in Finland, the Solway area, and in North America where they were decapitated and truncated and carved into spits and bars by the waters of lakes Algonquin and Iroquois[132] (see p. 474) and cut by outlet streams.[133] They underlie lateglacial marine clays,[134] like the Champlain and Yoldia Clays and those of the 100-ft beach sea of Scotland.

Small channels on the flank, more rarely on the crest, and some of the terracing of Swiss and German drumlins[135] may be lateglacial modifications by flood waters or osar streams.[136] Sands on the south side of drumlins occurred west of Lough Neagh in Ireland and in Ontario[137] where drumlins emerged from stagnant ice and the sun warmed the ground and melt-waters deposited the sands.

Ice-moulded. Drumlins, therefore, are not residual but original. Moulding by overriding ice is borne out by their regular smooth contours, readily distinguishable from the sharp scarps of water-sculpture; by their elongation—equiaxial domes are rare; and by their parallelism with subglacial stream courses,[138] with larger boulders and smaller stones[139]—the preferred orientation of pebbles in the surface layers shows that movement continued until the formation of the drumlin was complete. Drumlins are also parallel with lines of ice-flow, almost always that of the basal ice of the last glaciation.[140] The parallelism with the striae, detected first by Hall[141] and after a long interval by Close,[142] has now been observed countless times[143]; it is recognised in De Geer's term "radial moraines"[144] and is well depicted in the area of the Oder Lobe (see p. 1163), in north Ireland,[145] in the Michigan Lobe[146] and in Penck's maps[147] of the radially arranged drumlins in the glacier-lobes of the Alpine glaciation. Flow-direction is indeed better given by drumlins than by striae[148] which seldom continue for more than a few yards and are apt to be influenced by detailed topography. If relief altered the direction, drumlins show similar departures.[149] Where striae and drumlins do not coincide in direction, as occasionally noted,[150] this is due to a difference in age,[151] to a local deflection of the basal ice-currents[152]—these may cause the longer axis of the drumlin to be slightly curved[153]—or to a tendency of rock-drums to follow the strike of the rocks.[154]

Size and shape are alike related to the ice-motion. Drumlins in a particular area are apt to be of a particular size (see above) because the velocity hereabouts was the same; they were smaller near the end-moraines where the flow diminished.[155] The broader and shorter drumlins, often found on rising ground, were probably fashioned in more sluggish ice, the longer and more tapering ones in more active flow.[156]

W. B. Wright[157] ascribed the varying shapes of the Donegal drumlins to changes in the direction of flow. Deposition along the median line until the ice melted gave crested mounds; smooth, rounded outlines arose if deposition ceased before ice-motion. Oval shapes indicate alteration in the flow-direction, elongated forms consistency in direction. Sudden changes produced triangular, crescentic or forked plans. "Double deckers" or two-tier drumlins in the Genessee valley may also be related to such a change.[158] Re-advancing ice has re-shaped the drumlins and has added one or two distinct tails.[159]

Asymmetrical longitudinal profiles are likewise consistent with ice-moulding. The iceward end is generally broader, blunter, steeper and higher, the distal end, narrower and more tapering—"tadpole drumlins"[160] are extreme forms. This asymmetry, which Hall and Close early pointed out,[161] has

since been found in various parts of the world,[162] e.g. British Isles, Germany, Sweden, Switzerland and North America: even rock-drumlins display it.[163] Occasional statements to the contrary[164] apparently concern mounds which are not true drumlins or rest on erroneous flow-determinations.[165]

This asymmetry harmonises with physical law. Drumlins, as experimental evidence confirms,[166] are stream-lined[167]; they present their steeper face to the moving medium, in order to offer the minimum resistance to the flow by hindering the formation of vortices in the rear (which act as a drag on the moving body). Similar adjustment is seen in snow forms,[168] in certain dunes (Ger. *Strickdünen*) and sandbanks in rivers,[169] in the shape of torpedoes or fishes, or better still in those rheotrop animals, with hemistreamline form, which cling to the bottom of brooks[170] and because of lower velocity have less steep fronts and less extended tails—in these forms the average position of greatest transverse section is at a distance of about 36% of the total length from the snout. The streamlined form or rock-nucleus (see below) may keep the drumlin stationary, though erosion at the proximal end and deposition at the distal end may cause it to travel downstream (see p. 398).

Drumlins are normally laterally symmetrical unless they rest on slopes which rise with their minor axes. Occasionally, in a whole field one side (usually the eastern) is steeper,[171] the angle differing by 10° or 18°. The suggestion that this is due to erosion by postglacial streams[172] is erroneous and inconsistent with their form. It may be referable to a difference in the lateral pressure on the two sides,[173] connected possibly with Baer's Law[174] (see p. 36), but more probably with a change in the flow-direction during melting[175] (this sometimes added tails aligned in the new direction) connected with a readvance,[176] or a re-orientation that was arrested before the drumlin could be completely swung round.[177] Non-coincidence of drumlin trend with the direction of the boulders in the till suggests such a change in flow-direction.[178]

Ice-moulding is also seen in the pressure effects sometimes observed at the proximal end,[179] in the faint and parallel flutings along the flanks[180] and in the "washboard fluting" of low relief, as in New York State[181]; it is clearly irreconcilable with the superglacial origin occasionally advocated.[182]

The main facts concerning the structure of drumlins are generally admitted, as are the inferences that the shape is original and connected in some way with the depth, consistence and rate of motion of the ice. Here, however, unanimity ends.

An appeal to modern glaciers is unavailing, since drumlins *in statu nascendi* are unknown though accumulations somewhat resembling them have been described from time to time. An ice-moulded mound, 50–100 m broad and 200 m long, which simulated a drumlin in its composition and asymmetry, was discovered within the moraine of the advance of the Glarus Glacier of 1853.[183] Similar forms were afterwards detected[184] in connexion with the Steinen Glacier and Grübelferner and with the *Zungenbecken* of the Biferten Glacier whose mound was 230 m long, 120 m broad, and had a fluted surface and the composition of a ground-moraine. They have also been seen in Greenland[185] while drumlinoid curves[186] have been observed in the ice of Greenland and Spitsbergen (see below).

Submarginal. Drumlins may have been deposited by prolonged and powerful ice-modelling under the thick ice of maximum glaciation.[187] Thus rock-drumlins are parallel to and frequently indistinguishable from typical

drift drumlins; drumlins, as in the Vale of Eden, sometimes dispose themselves in accord with ice-fronts reconstructed by frontal deposits and marginal drainage; and the broad sweeping curves of the drumlin streams over wide regions date from a time when the ice was thick and detailed relief scarcely guided the flow. Drumlins, indeed, have been related in places, e.g. in Edenside and the Lough Erne and Lough Neagh regions, to basal icesheds (see below). The absence of moraines or other retreat features from drumlin areas (see below) and the discordant trend of the drumlins with the last ice-flow which seldom influenced their trend in any appreciable way, have been explained by supposing the ice to have been clean and sluggish or even stagnant at the dissolution.[188]

Some hold that drumlins were left behind at the last withdrawal[189] and were laid down submarginally[190] under active ice possibly 90–180 m thick[191] and 1–5 or 9 miles (8–14·5 km) back from the edge[192] as Wisconsin observations suggest.[193] Thus they are radially disposed to the retreat moraines,[194] are parallel with the late system of striae,[195] occur in successive belts, each a number of miles or kilometres in the rear of a moraine,[196] and are rare within 10 miles (16 km) of the outermost limit.[197] They have been linked with oscillations, each successive advance, within a major movement, adding one of the concentric layers.[198]

Drumlins are almost all restricted to the last drift,[199] e.g. the Wisconsin in North America or the Würm in Switzerland, though a few have been found outside it in south-west Ireland,[200] of Riss age in the Danubian portion of the Rhine ice,[201] of Illinoian age in North America,[202] and of the penultimate glaciation in South America.[203]

They were built up near the end of relatively thin ice which was moving slowly yet pressing sufficiently to mould the material. They were completed by the time the ice uncovered the area; their outlines are beautifully shaped; they underlie osar and moraines in the few cases where these are found together (see below); and they are occasionally trenched by osar streams or emerge from outwash or pitted plains.[204]

A submarginal origin is also suggested by the mutual exclusion of drumlins and such marginal or submarginal accumulations as osar and moraines.[205] Their absence from coastal Norrbotten has been linked with the floating of the ice during the retreat.[206] Högbom[207] thought drumlins were formed if basal melting was excessive, moraines where superficial melting predominated. The submarginal origin has the following support: drumlin trends within and without a marginal moraine are occasionally discordant[208]; moraines and drumlin belts are sometimes parallel (see above); chains of drumlin mounds may resemble moraines, as on the plain of St. Lawrence[209]; and a distinction from recessional moraines is sometimes difficult to make[210]—W. Upham's "Madison type",[211] with its large fluvioglacial core, is apparently part of such a moraine.[212] Its difficulties are arguments favouring the alternative view set out below.

While some glacialists think drumlins were built up principally of englacial grist,[213] others regard them as subglacial because their till is compact and mostly local[214] (distant material characterises the Winnipeg drumlins[215]) and their boulders are striated and increase in number toward the proximal end.[216] Others interpret the lower till as subglacial and the superficial layer of looser texture and more angular boulders as englacial[217] (see p. 383). The height of the drumlins may reflect the upper limit of the englacial debris.[218]

Erosion hypothesis. The thesis has been often advanced that drumlins are earlier drift sculptured by advancing ice or the waning ice-flood[219] ("subglacial exaration-landscape"[220]). They are end-forms or roches moutonnées in drift,[221] arched forms of least resistance stable beneath ice which is subject to a certain rhythm.[222] They are the outcome of the overriding or channelling of earlier drift sheets,[223] of terminal or median moraines,[224] of outwash sands and gravels,[225] of lacustrine deltas,[226] of osar and interglacial lake-clays.[227] Near Mount Billing and in Västergotland, they pass into osar and join up into corresponding lines,[228] and in Bavaria they are sometimes concentrated along a zone which coincides in orientation and width with the lines of median moraines.[229]

The reasons which prompt the idea of derivation from overridden moraines are the tendency of drumlins to group themselves in rows and belts suggestive of morainic lines[230] or within readvance moraines[231]; the fluvioglacial content of some[232]; the gradual passage from marked drumlins into morainic terrain[233]; and the drumlinoid forms, aligned with the ice-flow, in moraines[234] and connected with streamlines.[235] But, as critics point out, the material is homogeneous and compact, stratification and bedding are generally lacking, and water-worn detritus is insufficient or absent—the contention[236] that moraines are not always stratified and sections in drumlins rarely deep scarcely meets this objection. Moreover, although moraines overridden during readvances are known,[237] they nowhere assume the drumlin shape.

Accretion hypothesis. The seemingly opposite thesis, that drumlins are constructional, accreted on gentle slopes by successive additions from englacial or subglacial debris under a thin or weak ice-border, has been widely held.[238] Drumlins were built up, on the analogy of longitudinal sand banks in streams, where supply was excessive.[239] They grew at the margins of slowly moving masses of ice where the gradient was low (and subglacial waters were stagnant[240]) and the velocity was retarded by (*a*) local overloading[241] with thick preglacial soils, weak or soft rocks, or earlier drifts; by (*b*) fanning, as in radiating ice-lobes[242] and at the mouths of valleys or passes, so that the debris was pressed up along basal or longitudinal crevasses[243] or fell into them from above,[244] even if the stresses did not actually achieve an opening but merely weakened certain places[245]; and by (*c*) the ice passing into the sea[246] or on to rising ground,[247] as on the Vaudois plain, North German Plain, in New York State and British Columbia or around the Alpine *Zungenbecken*. In opposition to this view, it has occasionally been thought that the drumlins were formed at places of congestion.[248]

Drumlins, it has been surmised, grew along the plane of contact of two media (ice and ground), irrespective of any obstacle which was the disturbing exception rather than the rule.[249] The immediate cause may have been a change in the direction of pressure[250]; a rocky boss or obstruction,[251] such as a spur projecting from the margin of a *Zungenbecken* which the ice had sought unsuccessfully to remove by abrasion and plucking; a mass of earlier drift; or ice excessively charged with debris,[252] a lodgment of boulders,[253] a knob of frozen till,[254] or accumulations in glacier-stream channels.[255] An origin in surface crevasses is irreconcilable with the ground-moraine nature of the drumlin's material and their failure to penetrate to the sole. Genesis in basal crevasses or in subglacial drainage channels (see p. 253) does not harmonise with the abundance or shape of the drumlins.[256]

E. Ebers,[257] from studies in the Inn valley, has suggested the following

evolutionary stages: (*a*) striped arrangement of the ground-moraine; (*b*) incipient individualisation of the strips into irregular hills; (*c*) construction of steep, proximal end; and (*d*) full development.

American geologists, from observations on Greenland glaciers,[258] have formulated the laws of structure and behaviour of the ice where it encounters obstructions and develops thrust planes. G. Slater,[259] extending the laws, has shown that glacial disturbances are of two types; (*a*) a roche moutonnée type, associated with forward movement in which the ice, superimposed on the "core" dips off in opposite directions, accompanied on the iceward side by thrust planes and squeezed anticlines or flow folds, on the lee side by tip structure and flow curves taking the form of drawn-out lenticles; (*b*) a stagnant glacier type in which thrusts or imbricate structure occur throughout and anticlines in the lee. The structures are preserved as "glacial pseudomorphs" or fossil glacier structures, such as C. P. Berkey and J. E. Hyde described from North America[260] and K. Richter from north Germany.[261] Thus in Slater's opinion, the moutonnée cores of boulder-clay led to local pressure gradients and a plastering of additional material on to the sides of cores, the movements being along thrust planes—similar structures in Illinois have been attributed to freezing and thawing in frozen ground which was outside the ice-sheet.[262]

In west Greenland, layers of ice charged with basal debris were observed flowing over a rock-boss in a curve resembling the drumlin's longitudinal profile.[263] This true drumlinoid curve represents the balance between the force of overthrust and the tendency to accumulate material below the drumlinoid line of shear where wave-shearing,[264] possibly allied to Helmholtz's waves[265] or T. Rehbock's *Wasserwalzen*,[266] relieved the strains. Pressure and frictional melting against obstructions favoured deposition.[267] Experiments[268] show wax passing in like manner over low prominences and its lamination.

After the initial drumlin stage and the deposition of Chamberlin's "tails" and "pre-crag" deposits,[269] the ice dropped its load in the *morte-espace* in the lee,[270] this tendency being favoured by the strains that the drumlin nucleus set up in the ice.[271] Pressure compacted and moulded the debris of these ramps as in Case's experiments.

While those drumlins which have a rock core were presumably stationary, those without may have migrated as the upstream side was eroded and the material was transferred downstream[272]—Nova Scotia drumlins have migrated from the slates on to the quartzite country (see p. 379).

W. Upham expressed a view,[273] more or less similar to that of DeLuc,[274] that drumlins were formed when thin, debris-laden ice was overridden during a temporary thickening of an advance; differential shearing incorporated the drift in great lenticles.

Accretion and moulding. The preceding observations establish the reality and importance of accretion in building up drumlins. The activity of the ice during this time is demonstrated by the following observations: drift drumlins pass into rock-drumlins[275]; rock emerges at the proximal end[276]; the flanks are fluted[277]; crescentic grooves sometimes encircle the iceward end[278]; and ice produces drumlinoid forms.[279] Accretion, however, was opposed by erosion which prevented the drumlins from exceeding a certain uniform elevation in any particular region.[280] The control was complex. If suitable material was inadequate, the mounds were either isolated and sparse or

irregularly fashioned, small or embryonic[281]; drift-veneered knobs likewise tended to be small.[282] The size depended too upon the consistency of the drift. Adhesive clay, by acting as a lubricant and adhering to a mound already started, facilitated the building up process[283]; drumlins in Nova Scotia are restricted to slate belts (see p. 379) and the boldest and most typical ones in New York rise from clayey rock, e.g. Salina Shales.[284] If made of sand, on the other hand, they are low[285] as in Switzerland, Bavaria and about Dorpat, or less perfect,[286] as in the Vale of Eden. Drumlins are relatively broad in granitic areas[287] or where the drift is thick,[288] but are long and thin where the drift is thin,[289] and are poor or absent if material is friable or bouldery[290] as in Chamberlin's "immature nuclei" or "till tumuli".[291] The size was also governed by the depth and rate of flow of the ice[292]; the mounds were apt to be more or less rounded if the flow was variable or indefinite[293] but were attenuated[294] where the ice moved more quickly or rigidly and consequently more erosively. They were high and broad if situated among roches moutonnées and low if the relief was low and coincident almost with glide planes in the ice, as around the Gulf of Bothnia.[295] They may belong to the intermediate region where erosion and deposition tended to be in conflict.[296]

Conclusions. Drumlins are an expression of the equilibrium[297] between the erosive action of ice and the opposing forces of the solidity and cohesion of the material; the "plastering on" and "rubbing down" processes[298] took place at the same time or in rhythmic alternation. The ice moulded the accumulations which were limited in their height and other dimensions by its erosion. Hence, the two views of erosion and accretion are reconcilable; they emphasise different aspects of one process; in some cases the one was dominant, as in the case of the rock-drumlins, in other cases the opposing process. In certain regions, the processes seem to have been at work at the height of glaciation, in others during the retreat.

Much, however, remains to be elucidated. This is inevitable where so many factors are necessarily indeterminate, such as the rate of flow and forward thrust of the ice, the thickness, pressure and plasticity at the place of drumlin formation, and the amount of englacial material at each stage of glaciation in the various parts of the ice-sheet. The problem of when the ice formed till plains or "non-drumlin areas" and when drumlin landscapes is still unsolved. Drumlins may occur if there were minor readvances[299] (see p. 397) or overloading, plains where the drumlins were later destroyed[300] (most improbable). Alternatively, drumlins grew on a country of uneven bedrock and plains arose where the subjacent rock was smooth and the basal drift was evenly distributed and the pressure at the base of the ice was uniform,[301] or where the velocity did not reach a certain critical velocity, as in the centre of ice-flow diamonds[302] (see p. 715). Further progress in the understanding of these and other factors at the base of the ice will depend on the further elucidation of the basal ice-flow.

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Chapter XX

MORAINES

Classification. The rarity or absence of exposed rock and the low velocity and slight erosive power of the ice cause debris to be scarce or lacking on the surface of sheet-ice or plateau glaciers. L. Agassiz[1] early postulated this for the Pleistocene glaciers of the Alps. Confirmation came later from observations on modern ice-sheets[2] in Greenland and Antarctica and on the plateau glaciers where superficial moraines are few, short and narrow; dust particles in the air over Ross Barrier are only half as many as over the mid-Pacific.[3]

Debris, however, is usually plentiful around the margins of ice-sheets and glaciers. This material, designated moraine, is named lateral when distributed along the sides, and frontal or terminal if laid down at the front or end. Inner lateral moraines unite into a median moraine at the confluence of two glaciers and travel down the united glacier along the plane of contact.

These terms were introduced by Agassiz[4] in 1838 and independently by Godeffroy[5] two years later, though Hugi[6] had already recognised the various types in 1830 from which time morainic studies may be said to have assumed a scientific aspect. In de Saussure's day, the word moraine was variously and wrongly spelt[7]; H. Besson,[8] for example, in 1780 wrote *marême*. De Saussure gave literary sanction to the Savoyard word *moraine*, derived apparently from the Latin *morena*, a bank of stones. Some glacialists, including such early writers as Besson, Charpentier and de Saussure, restricted the term to deposited moraines, while later glacialists like Agassiz, Forel, Heim and Penck distinguished between moving and deposited moraines. The distinction was recognised in the classification drawn up by the International Glacier Commission which, at a conference on the Rhône Glacier in 1899,[9] attempted to remove the confusion in the nomenclature that then existed. The Commission's classification is subjoined.

Classification of Moraines

Moraines mouvantes (Moving moraines)	Moraines superficielles (Surface moraines)	Moraines latérales (Lateral moraines) Moraines médianes (Median moraines)	
	Moraines internes (Internal moraines) Moraines inférieures (Basal moraines)		
Moraines déposités (Deposited moraines)	Moraines rempart (Dumped moraines)	Moraines longitudinales (Longitudinal moraines) Moraines marginales (Border moraines)	M. riveraines (Flank moraines) M. frontales (Terminal moraines)
	Moraines de fond (Ground moraines)	Moraines profondes (Ground moraines) (Drumlins)	

The classifications of P. A. Øyen[10] and A. v. Böhm[11] involve new terminologies and an unnecessary differentiation of subtypes.

The following brief description does not stress the difference between moraines at rest and those which are not; for these differ in no essential respect. The basal or ground moraines have been the subject of the two preceding chapters.

Lateral moraines. Superficial moraines (E. de Billy's "dorsal moraines"[12]) include lateral, median and terminal moraines. Of these, the first are perhaps the simplest in composition and mode of occurrence. They form sharp crested ridges and usually show no arrangement according to size or weight.[13] They are derived from several sources, viz. subaerial ablation in the firn, heat and cold acting upon the adjacent cliffs, streams and torrents fed by rain storms in wet weather, and erosion of the glacier's bed and sides.[14] Frost acts along the structural planes in the bare walls and builds scree-chute gullies. Its operation is supplemented by rock-falls and by avalanches which pour down couloirs and contribute much angular material (see p. 579).

The factors which control a moraine's size are many. They include the rate of motion of the ice; the inclination of the valley wall and the presence of structural benches which determine whether the detritus shall rest on ice or on rock, gentle slopes contributing little and being in some degree protected from further action by their cover[15]—the effect is seen, for instance, in the contrasted size of the moraines east and west of Lago di Garda; the nature of the rock in the adjacent crags—large moraines skirt the base of soft tuffs and breccias in Iceland[16]; aspect,[17] as for example in the eastern Alps and Dakota; and landslips which are prone to occur if there are oversteepened sides or suitable structures. The "1850 moraines" of the Alps are up to 100 m high and moraines in Greenland up to 200 m high. Continuous moraines may be fed at only one place and receive additions at no others. The rocks may, however, fall on to points at unequal distances from the ice-edge[18] (Charpentier's *moraines doubles, triples et multiples*[19]).

Glaciers in the Pamirs and Himalayas (except in the north and east) are noteworthy for their extensive morainic cover[20]; the cliffs are steep and high, heat and cold are extreme, and avalanches, some of them shaken down by earthquakes, are highly frequent.[21] Dust blown from Baluchistan and Persia has added its quota.[22] Occasionally the moraines are 300 m high.[23] The larger glaciers of the east side of the South Alps of New Zealand are completely covered with debris for the last few miles of their courses.[24] Pleistocene moraines in the Hohe Tatra owed their unusual size to jointing in the granite and to high relief.[25]

Superglacial material is generally without polishing, scratches or other signs of glacial action.[26] Such englacial and subglacial constituents in the moraines as do occur are recognised among the angular fragments by their subangularity[27] and occasionally and especially in limestone districts by their smoothed and striated boulders, though these are strikingly rare in association with modern ice.[28] Till however does occur in modern moraines, e.g. in east Greenland.[29] Moraines extruded along the borders of nunataks, particularly on the impact side (*moraines par obstacles*[30]), e.g. on the Jensen and Dalager Nunataks[31] and on others in Greenland[32] as well as in Spitsbergen,[33] consist largely of such ground moraine.[34] This is due to increased frictional resistance at the base of the nunatak as Tyndall surmised and Chamberlin proved.

The "crag and collar" or "crag and cap" moraines of Pleistocene glaciers[35] were comparable.

Lateral and subglacial streams and the "tabling" of surface debris (see p. 60) contribute rounded material.

Lateral moraines often strongly resemble landslips topographically,[36] notably if they have settled or shifted since the ice withdrew or erosion has dissected the slips. The distinguishing criteria are (*a*) the relationship to the rocks *in situ*; (*b*) the condition and nature of the material—a moraine is heterogeneous, a landslip, local and uniform in its composition; and (*c*) the topography of the accumulation, e.g. a moraine is approximately even in elevation and continuous across a tributary valley (*morena insinuta*[37]) where a landslip ridge could not form, or links up with a terminal or recessional moraine by an intermediate ("longitudinal"[38]) moraine.

There may also be confusion with screes, unless the accumulation is wall-like and encloses glaciated boulders; with solifluxion ridges, as in the German Mittelgebirge[39]; with mounds of loose material formed by sliding in wet weather[40]; with eroded alluvial cones[41]; with lateglacial avalanche heaps, as in the Sareks region[42]; or with terraces, as exemplified on Gaussberg (371 m) where outcrops of hard lavas[43] have been interpreted as moraine.[44]

Median moraines. Composite glaciers carry on their backs one or more ribbons of debris whose serpentine courses follow the glacier's windings. De Saussure[45] expressed the view which persisted until much later[46] that the median moraine was thrust up by the ice moving from the glacier's side. The true explanation that it arises from the union of the inner lateral moraines at the point of confluence of two glaciers was given by B. F. Kuhn[47] in 1787 and much later by Agassiz and Charpentier who recognised that a median moraine may originate if only one of the glaciers has a lateral moraine.[48] It was reserved for Godeffoy[49] to give the simple formula connecting the number of median moraines with the number of confluent glaciers. On the unbranched glaciers of Sweden, median moraines are obviously very rare.[50] Median moraines also stream from the lee sides of nunataks[51] or the ends of spurs projecting from the sides of glaciers[52] or appear at the place of emergence of inner moraines.[53] They consist of scree or landslipped material; in the latter case, they form more or less separate heaps (*moraines passagères* or *moraines d'éboulement*,[54] *moraines circonscrites*[55]).

Lateral moraines gradually unite and mingle their materials, though they may retain their petrological differences[56]; one half of the moraine of the Aar Glacier is dark mica-schist, the other white granite.

Median moraines are not necessarily near the median line. Their position with respect to this depends upon the relative strengths of the confluent glaciers. If these are very unequal, the median moraine may be pushed directly across the mouth of the weaker glacier[57]; they constitute Agassiz's oblique moraines.[58] Since a glacier's components do not invariably reach the snout, a median moraine may become marginal if a lateral stream dies out[59] or two such moraines may unite if the central component ceases.[60] Festooned median moraines arise from a sudden advance of lateral glaciers[61]; there may be a series of them one within the other.

Median moraines, acting collectively like a glacier table, protect the underlying ice. Their ridges of ice, capped with moraine, widen as they travel by sliding from their flanks[62] as a result of the glacier's curvature and the slowing down of its movement. This spreading partly explains the dirty

end-portions of some glaciers, e.g. the Z'mutt and Aar glaciers of Switzerland and others in New Zealand. Occasionally, as in the arid regions of the Mount Everest part of the Himalayas, median moraines may repose in sunken troughs.[63]

A median moraine is composed not only of angular surface fragments but of subglacial material,[64] as C. de Seue noted in Norway, F. Simony on the Dachstein glaciers, Penck on the Sonnblick glaciers, and S. Finsterwalder on the Vernagtferner. It also contains detritus rounded and glaciated along the plane of contact of confluent glaciers which move unequally,[65] as on the Fürkelferner where vertical boulders, disposed parallel with the flow, lay in numerous parallel lines. Penck[66] explained this phenomenon on the Wurtenkees by supposing that the seam opened occasionally to admit boulders which reappeared by melting. That median moraines may extend downwards to bedrock was suggested by E. de Martonne[67] and has been proved in a subglacial tunnel 500 m from the snout of a Greenland glacier.[68]

If a median moraine arrives at the snout, it helps to build the terminal moraine. If it is laid down before doing so, as at the final melting, it may retain its linear shape (pl. XIIIB, facing p. 416).

Median moraines of Pleistocene age are relatively rare[69]; they were either destroyed by lateglacial or postglacial streams or masked by accumulations which were let down at the dissolution.

Inner moraines. Median moraines which, as Forbes[70] observed, gradually appear at the surface without any visible rock-source, spring from emerging inner moraines. These vertical and parallel bands of debris, seen in the walls of crevasses[71] or the terminal faces of floating glaciers, e.g. the Barry Glacier of Alaska, are derived from avalanche cones[72] or from fragments engulfed in crevasses[73] or plucked from the summits or lee sides of rocks jutting into the ice.[74] Ground moraine may also be squeezed up along a plane of contact, as actually observed,[75] or demonstrated by marine shells in some Icelandic median moraines.[76] Ridges of debris transverse to the flow[77] (*Quermoräne*, cross, oblique, melt or pseudo-terminal moraines) may come from basal crevasses[78] or steps in ice-falls.[79] The closely spaced "ice-crack" moraines of Canada[80] ("washboard moraines"), which were formed parallel to the ice-margin during the closing stages of glaciation, may be "annual moraines" (see p. 1151) or related to these cross moraines. Finsterwalder's mathematical theory (see p. 121) explains how inner moraines appear where the streamlines emerge that originated in a union of glaciers in the firn region.

The inner moraine, which springs from subglacial sources and, as in the plateau glacier type, has some glacial material,[81] broadens downwards as on the Hochjochferner.[82] This partially explains the wider spread of debris at the end of many composite glaciers like the Aletsch, Lower Aar and Gorner glaciers of the Alps and many in the Karakoram–Himalayas where the whole breadth of the glacier is covered for many kilometres from the ends. In these mountains, however, the great spread of material was derived from almost ceaseless avalanche falls: the Zemu Glacier lies under such moraines for 18 km from its end, the Baltoro Glacier for 50 km. In some cases, it is ground moraine brought to the surface along shear-planes (see p. 117).

Terminal moraines. Terminal moraines are of many kinds. The simplest is the snow-slope moraine, variously designated firn moraine,[83]

Schneeblockwall,[84] *Schneeschuttwall*,[85] "nivation ridge",[86] "winter-talus ridge"[87] or "protalus rampart".[88] These fringes of talus, which may reach the dimensions of lofty ramparts, are disposed in irregular scallops far out from cirque walls or in the shadow of escarpments or steps beyond the limits of rolling of modern scree. They were erected, as is observed at the present day,[89] at the foot of perennial or winter snow-wreaths which, often aided by avalanche falls, rose against the rock-face and filled the hollow now frequently occupied by small tarns.[90] Terracing results from seasonal variations of growth. The moraines consist of angular scree which was shattered from the crumbling heights above the snow and has glissaded to the bottom. Transport and solifluxion facilitate more rapid weathering than on similar faces free from snow. Rock slides have contributed to the accumulations.[91]

Pleistocene moraines of this kind were widespread,[92] as in the Alps, German Mittelgebirge and British Isles; their short distance from the cliff, their regular form, and their lack of striated boulders or of drainage gaps are diagnostic.[93] They may, however, pass into true moraines if the snowfields develop into small cliff glaciers[94] or into block moraines as in the Pyrenees.[95]

Morainic debris concentrates at the ends of valley glaciers. Surface moraines, lateral and median, mix inextricably with the emerging englacial detritus as the plane of ablation descends. By protecting increasingly the underlying glacier, they retard its melting and blend it imperceptibly with its moraine. This behaviour in arctic and temperate latitudes is reversed in the Antarctic[96] where surface fragments become englacial by countersinking and by the outward drift of surface snows (pl. XIVA, facing p. 417).

Terminal moraines, which are looped across valleys or over plains, contain therefore angular superglacial material, with a little englacial and subglacial material as Besson (1777) and de Saussure (1779) early observed in the Alps and later investigators[97] recorded both here and in the Arctic and in Pleistocene moraines. Though usually unstratified and unassorted,[98] they sometimes harbour waterworn and stratified sand and gravel,[99] where melt-water is copious, and consist entirely of well-washed gravels and sands where the ice stands in water.[100] Cirque moraines which are coarser and steep fronted have particularly angular material, as Collomb[101] emphasised; their fragments are homogeneous and lack the finer elements. Lateral moraines in the Swiss Alps likewise have much angular debris because their valley walls are steep and high.[102]

Block moraines (*Geschiebewälle*[103]; "bear-den moraines"[104]) occur if the rocks are well jointed. Their blocks are scattered over many acres with little or no fine material. Monthey in Switzerland furnishes the classic example,[105] the moraine being 3 km long and 300 m broad. Other instances[106] have been observed in Canada and north Germany and may be due to the wasting away of the finer material or, more commonly, to boulders rolling down steep frontal slopes of the ice.[107]

A terminal moraine is generally related in size to its valley though in some cases it is enlarged by a core of stagnant ice.[108] It is especially big in the "morainic amphitheatres" (*anfiteatro morenico*) of B. Gastaldi[109] where the material was greatly concentrated. The moraines sweep in vast arcs around the southern ends of the north Italian lakes,[110] as at Ivrea—the most spectacular moraine in Europe—and around Lago di Garda, rising 500 m above the lake, compelling roads and railways to adopt sinuous curves or to burrow in tunnels through them. In Iceland, they rise up to 110 m above the sur-

rounding plain[111] and in South America to 400 or 600 m above lake-surfaces.[112] Distinctness and size are a function of many variables, the chief of which are the supply of debris, the duration of the halt, and the state of the glacial drainage. Moraines, however, do not accumulate as a matter of course but only if conditions are favourable. They may be poor or wanting if glacial flood waters wash the material away,[113] e.g. below the Fedchenko Glacier of the Pamirs, the Tasman Glacier of New Zealand or the Pleistocene ice in the Lake District, or the seaward side of the coastal mountains of Alaska, or if they are buried under outwash sand or lake-deposits[114] or later boulder-clay.[115] There may also be none if the ice is clean or wastes downwards near its margin[116]; if it advances and then without pause immediately shrinks back; if it retreats steadily and drops its load uniformly as a superficial cover to the rock or ground moraine; or if advances push the accumulations to one side as a stau lateral moraine.[117] Drift borders without end moraines have long been recognised and described as attenuated (see p. 899).

Moraines may mark pauses both at the limit of glaciation (Chamberlin[118] would restrict the term terminal moraine to these) and during recessions (Chamberlin's "peripheral moraines"; C. King's "recession" or "recessional" moraines[119]). Some of the most marked are the product of a rapid readvance following periods of stabilisation, though some of these readvance moraines are known by their thickness rather than by any distinctive topography.[120] They may be a response to a fall of temperature or to increased nourishment, and are lacking in continuous retreat as Charpentier[121] noticed or if the edge oscillates widely or advances or retreats without becoming stationary. They generally require active motion up to the margin. Sometimes, however, as in the case of the "ablation moraines"[122] of R. S. Tarr (G. Hoppe suggests the term "hummocky moraine") they may mark the emergence of inner moraines and sweep in crescentic lines of chaotic appearance parallel with the ice-margin on the outcrops of shear-planes on dirt-laden layers, as on the stagnant Variegated Glacier of Alaska and in Iceland. They are indistinguishable from ordinary recessional moraines when the ice finally melts and the detritus is let down.[123] Ablation moraines, as in modern Alaska,[124] may give rise by differential melting to a depression or interior flat above the outer bulb underlain by clear ice (pl. XIVB, facing p. 417).

Recessional moraines may be independent of climatic factors if the marginal ice is congested with englacial material and retarded.[125] It has been urged from the experiments of Tarr and Engeln[126] that continuous precipitation may make the flow discontinuous at the margin, or that excessive melting by steepening the slope might rejuvenate the flow and initiate an advance.[127]

The terminal moraine's position may be controlled by rock-obstacles which resist progress by the ice[128]: this view has been expressed (with doubtful validity) for the Baltic Moraine[129] where such lobes as those of Lübeck, Kiel and Flensburg were controlled by the preglacial relief.[130] Moraines are often found at the mouths of valleys, e.g. the Gschnitz moraines of Tyrol,[131] or just above the points of confluence with tributaries.[132] The first position may be determined by tides and currents if the ice enters a sea[133] or by deployment, which by expanding and thinning the ice makes ablation exceed the supply—this equilibrium has been associated with the occurrence of the thresholds at the mouths of fjords.[134] The second position may be governed by a change of gradient and better drainage which reduce solifluxion[135] or by depriving a glacier of a feeder as it retreats to a new position of equilibrium.[136]

Moraines which consist of ground moraine and have been built forward and upward so as to convert the glacial troughs behind them into hanging valleys have been termed "pedestal moraines".[137]

Boulder-belts. Erratic boulders are sometimes densely concentrated into strips of country transverse to the flow. They are either scattered thickly over the ground or piled up as boulder-walls, up to 100 m broad. Such boulder-belts occur in north Germany[138] (*Geschiebestreifung*), notably within the Baltic Ridge and close to its southern edge, as well as in parts of the United States.[139] Though they may be merely ground moraine with an unusual abundance of erratics[140] or arise from the washing of boulder-clay,[141] they are more probably of the nature of end moraines,[142] belonging to the subtypes of dump[143] or stau-moraine[144] (see below), since they run parallel with retreat moraines and have more angular and distant erratics than the local boulder-clay.[145]

Push moraines. Chamberlin[146] distinguished several morainic subtypes: (1) "dump moraine", composed mainly of superglacial and englacial material dumped at the front of the ice; (2) "lodge" or "submarginal" moraine, consisting of subglacial debris lodged under a thin ice-edge and passing into "till billows" and subject to the mechanical action of oscillating

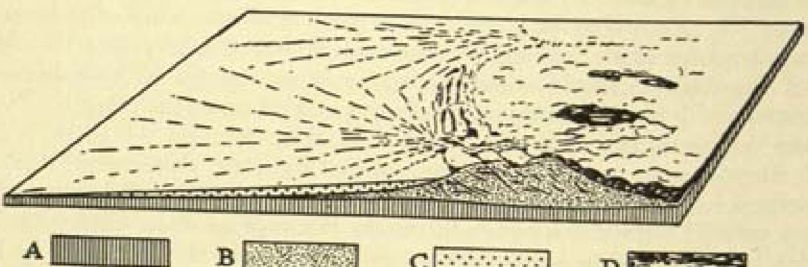

FIG. 73.—Diagram of a push moraine. A, preglacial soil; B, pushed layers, mostly of preglacial age; C, fluvioglacial sediments; D, ground moraine (boulder-clay). C. H. Edelman, *1880*, p. 15, fig. 6.

and oversliding ice—they may be a few miles broad and, according to Chamberlin, the predominant type in North America; and (3) "push moraine" (other terms are "shoved" moraine,[147] *Stau-*[148] or *Stauchmoräne*[149]; Dut. *Stuwwal*) which had two varieties, the one composed of glacial material, the other of local and non-glacial debris which the advancing ice ridged up in its path. The lodge and dump elements had long been recognised.[150]

Push moraines, first recognised as a type in the 70s of last century,[151] are usually broad, smooth and massive. They are relatively free from small surface irregularities, transversely asymmetrical, and frequently arc-shaped[152] (Fig. 73). In modern Spitsbergen, they are up to 80 m high and arranged in parallel rows *en échelon*.[153] Their composition largely depends upon the nature of the ground in front of the advancing ice; they may consist of subjacent rock, of outwash sand and gravel, of marine sediment, of banded clays and silts from glacier-lakes or of ground moraine. In this case, as in some accumulations in north Tyrol and south Bavaria,[154] topographic relationships only reveal the morainic origin. Arched structures and folds have often been

noticed,[155] e.g. in Prussia and in Sylt, as well as *Schuppen* and obliquely faulted strips in association with modern glaciers, e.g. in Spitsbergen.[156] Drumlinoid shapes may occur if the mounds have their axes perpendicular to the ice-face.[157]

Active push moraines lie in front of existing glaciers in Spitsbergen,[158] especially with marine sediments at the head of ice-fjords. In Greenland and Iceland they are less well displayed since ice-free territories with abundant unconsolidated detritus are rare and the crystalline rocks are less favourable than Spitsbergen's Mesozoic and Tertiary strata. Nevertheless, they have been occasionally reported from both these countries[159] as well as from Novaya Zemlya[160] and from North America's Illecillewaet and Asulkan glaciers.[161] They likewise characterise Alaska[162] where they have uprooted trees and pushed them over (see p. 141), pressed gravels into saddles and troughs, disturbed beaches, and fluted surfaces in the flow direction. Over-steepened bedding, brecciated strata, and folded and faulted structures are other tokens; the ice itself may be folded along its margin.[163]

A number of earlier as well as later geologists have held that all or most terminal moraines, as in Holland and north Germany, were of this origin[164]; push moraines have indeed been considered synonymous with terminal moraines.[165] True push moraines and effects of push have been observed in both Europe[166] and North America[167]: the Krefeld-Nijmegen moraine is an example.[168] But to extend the term to embrace all terminal moraines is obviously to exaggerate.

Stau-osar. A series of features, closely related to push moraines, are the *Stau-osar* (*Durchragungsmoräne*[169]; *Aufpressungsosar*[170]). Boulder-clays, e.g. in north Germany[171] and less commonly in Denmark[172] and Sweden (where their rarity is possibly to be linked with the thinness or absence of drift beneath the recessional deposits[173]), have been pressed up with arched bedding along lines parallel with the flow. The Lower Diluvium and even bed-rock, including Mesozoic and Tertiary strata in Holland, have been squeezed into the Upper Diluvium as arches or forced into it as apophyses[174]; according to Schroeder, two parallel depressions or weak negative downfolds accompany them. They may have arisen in the marginal zone where subglacial detritus was pushed up along subglacial drainage lines or radial crevasses,[175] possibly as the result of some movement of the ice. Some of the radial moraines of Sweden, e.g. in Västerbotten, Norrbotten and Dalarne, and of the Varanger peninsula, may have arisen in this way.[176] Observations on modern Icelandic glaciers show basal drift being forced up into radial crevasses.[177]

Finsterwalder's theory and moraines. Finsterwalder's mathematical theory of ice-flow (see p. 120) throws much light upon the origin and composition of the various moraines.[178] Rock-debris falling on the very margin of the firn passes along the floor as ground moraine and is finally ejected in glaciated form in lateral and terminal moraines around the glacier's edge. Rocks above the firn yield material which follows the lines of flow and emerges by ablation as median moraine down the tongue. Debris, however, which drops on the latter is carried superglacially. It does not penetrate the ice save along crevasses whence it reappears, e.g. below ice-falls, as Agassiz[179] noticed. Fragments from the lee of rock projections in the tongue extend as median moraines to the base and widen down the glacier by superficial melting. Transverse ridges or prominences rising into the ice drive up all the

streamlines: they carry up the ground moraine to form an englacial moraine which widens as the glacier descends.

The lack of moraines and their differential melt-forms in the firn, except at the base of precipitous cliffs, accords with the theory and with general observation[180]; the debris at the confluence of névés passes under the surface as inner moraine (*Adermoräne*[181]) to emerge at or below the firnline.[182] Firn moraines, to use the term of the brothers Schlagintweit[183] (Ratzel's firn moraines are snow-slope moraines) have been occasionally described[184] but appear to rest upon erroneous determinations of the local snowline or upon some irregularity in the floor of the firn basin. Corrie glaciers have virtually no lateral moraines.

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Chapter XXI

ESKERS

Terminology

The words *cam* and *kaim*, signifying in Celtic crooked or winding,[1] are used generally in connexion with rivers, less commonly with ridges. Esker (Irish *eiscir*; Welsh *esgair*, *escair*) and kame are etymologically equivalent, and accumulations bearing these names have often been regarded as glacially synonymous,[2] as by Close[3] who introduced the word esker into glacial literature. Others equate esker with ridges and kame with mounds[4] or seek to draw an unworkable distinction between kames less than 2 miles (*c.* 3 km) long and osar of greater length.[5] It is suggested[6] too that kame be reserved for water-laid marginal moraine and esker for accumulations in the channels of subglacial streams, a genetic distinction which resolves itself in part into elongation across or along the flow-direction. Since, however, Scottish kames[7] embrace true eskers as here defined and Irish eskers[8] true kames, confusion immediately arises. Hence, it appears advisable to use esker (*sensu lata*) as a general descriptive term for glacial sands and gravels arranged in mounds or ridges, irrespective of their genesis,[9] and confine the term kame-moraine[10] to accumulations which have the same significance as moraines and like them are parallel with the ice-edge. Osar, from the Swedish word *Åsar*, should be employed for deposits associated with subglacial tunnels which were generally transverse to the margin.

To interpret eskers correctly in their manifold forms is one of the most thorny of glaciological problems. We may recognise three main types: kame-terrace, kame-moraine and os. All as a rule were features of thin and feebly flowing or stagnant margins.

1. *Kame-terraces*

Kame-terraces[11] (variously named "moraine-terrace",[12] "lateral moraine-terrace",[13] "glacial lateral terrace"[14] or "ice-contact stream terrace"[15]) have often been discovered along the margins of both modern[16] and Pleistocene[17] glaciers—they were first correctly interpreted in the Scottish Highlands in 1860[18] and described from modern glacier-margins in Alaska in 1893.[19] Usually narrow and flat-topped, and pitted with kettles, they may skirt one or both sides of a valley for considerable distances or lie in the angle at the junction of valleys.[20]

The outer edge of their sands and gravels, where still preserved, is commonly a steep ice-contact slope (J. B. Woodworth[21] was apparently the first to use this term), the faithful cast of a former ragged ice-margin, affected by slumping and dissected by later streams. This dimpled and scalloped face—recently formed ice-contacts in Alaska have been described by R. S. Tarr[22]—may be faced with boulder-clay or stony lateral moraine[23] or surmounted by the push moraine of an oscillating ice-edge.[24] Some terraces build out antecedent divides into plateaux[25] or make shelves below sea-level if the ice passes into the sea.[26]

These terraces, which may display a seasonal periodicity,[27] were laid down by aggrading streams issuing from the ice or draining from the land and coursing inconsequently along a glacier's flank. They filled crevasses or subglacial hollows ("subglacially engorged eskers[28]) with water-laid sand and gravel which at the dissolution appeared as narrow ridges, winding and tapering towards the centre of the valley.[29] The terraces incline downstream, expand at the mouths of valleys into truncated fans, and in the case of the larger ones grade downstream into valley-trains. They are sometimes discontinuous, the segments being disposed at different heights between rock-bosses or spurs which mark the sites of falls or rapids. Successive terraces, resembling a giant staircase on a mountain side, mark as many halts in the melting down of the constraining barrier. Some of the terraces were formed in long, narrow marginal lakes and have no downstream gradient.

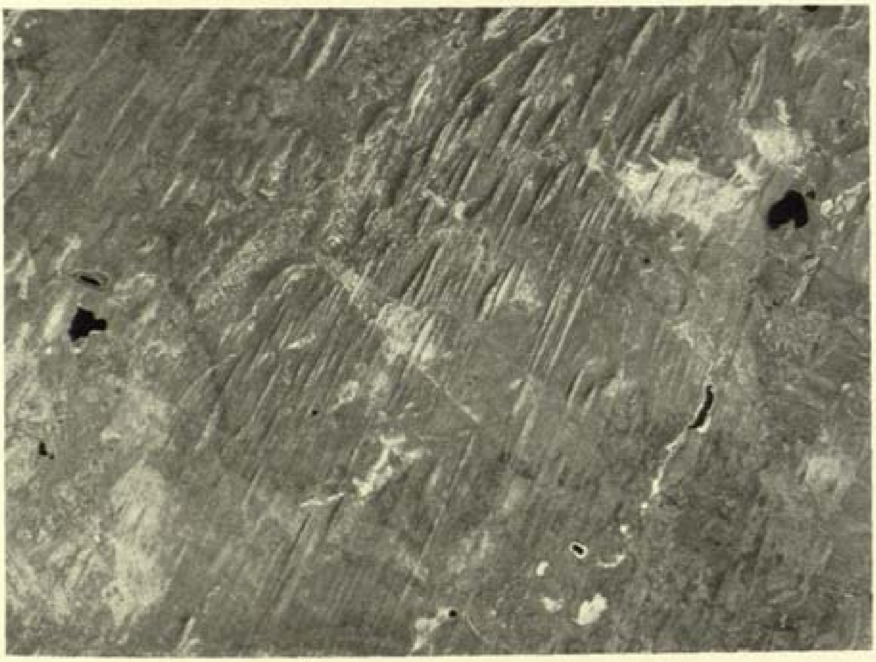
Kame-terraces may, on casual inspection, be easily confused with the strandlines of glacier-lakes; the mistake has not infrequently been made.[30] Their surface, however, is broader and less regular, they enclose morainic debris and they occasionally occur in positions and at heights that preclude lake-action[31] (pl. XVA, facing p. 432).

2. *Kame-moraines*

Topography. Kame-moraines present a striking and characteristic appearance in a landscape. Rapidly undulating and swelling like a choppy and billowy "sea" of confused hills, they form close complexes of tumultuous rises and hollows intertwined with parallel or tortuous ridges which unite and interlock in a most bewildering manner. The knolls sometimes appear so artificial that they have been mistaken for barrows, tumuli or Roman forts or interpreted as the dwellings of a fairy race.[32] The accumulations thin on to the hillsides and elsewhere either fade gently into unobtrusive plains or end in abrupt and sharply defined ice-contact slopes with strong cusps or deep re-entrants. The inner face, as in the case of deltas and fans, is an ice-contact slope; its steep and linear form is a mould of the ice-face modified by slumping after the ice withdrew (pl. XVB, facing p. 432).

Kame-moraines are dappled with countless closed hollows which are saucer or bowl shaped, oval or elliptical. These range in depth to 45 m[33] and in diameter from a few meters to over 1000 m (the Mecklenburg *Soll* averages probably 41 m across and 3·25 m deep[34]) or, as in Minnesota, up to 8 miles (13 km); few are more than 6–9 m deep. They often hold lakes, lakelets or ponds, commonly without visible outlet, if their bottom is below the local water-table or is floored with clay. Flat marshes, peat bogs or alluvial tracts mark their former sites. These kettle-holes—the name comes from the Kettle Range in south Wisconsin a few miles west of Lake Michigan[35]—are the *Sölle* of Mecklenburg,[36] *Pfühle* of Brandenburg,[37] *Sol*, *Sool* or *Soll* of Switzerland, and *Kaulen* of Posen. They are extremely numerous as is implied in W. Ramsay's term *Kleinseegebiet*; in Posen, they number possibly 30,000, in Mecklenburg, 35,000–40,000.[38] They are frequently markedly parallel with the kame-moraine[39] and in Bavaria are apparently deeper at the distal end,[40] though this observation has not been confirmed elsewhere.

Structure. Kame-moraines consist mainly of bedded and assorted sands and gravels, whose pebbles are well rounded and rarely striated, though the



A. Ispatinows, north of Weitzel Lake, Cree Lake area, Saskatchewan [Royal Canadian Air Force : Crown copyright]

B. Superglacial moraines on Logan Glacier with innumerable cirque glaciers about Mt. Logan (19,850) and King Peak (17,850), Alaska [H. Bradford Washburn]

(facing p. 416)

A. Terminal and lateral moraine, Loch Callater, Braemar, Aberdeenshire [Geol. Surv. Gt. Britain : Crown copyright]

B. Ablation moraine on stagnant snout of Chitina Glacier, Alaska, which is studded with kettle-holes and at the tip covered with trees over 100 years old [H. Bradford Washburn]

(facing p. 417)

rounding has often only partly altered the former faceted shapes of ice-worn grist. The stratification is undulatory or arched, irregularly lenticular or current-bedded, or inclined outwards to the margin. It is sometimes traversed by faults or local disturbances. The material is in places almost indistinguishable from ground-moraine[41] and may possess kame-form without kame-structure. The till portions, as along the sides of valleys, have smoother slopes and more softened outlines than the gravelly members which have steep and abrupt contours.

Distribution. The broad belts of the kame-moraines, often miles wide, characterise plains, wide open valleys, and submontane strips where the ice debouched upon open country. Commonly, as on the plains of north Germany and North America, they are festooned in loops with re-entrants usually buttressed upon higher ground and marked by a more pronounced topography. In this "intermediate" or "interlobate" type[42] assorted material from the important drainage lines of the glacial melt-waters is relatively large; the interlobate moraine between the Green Bay and Lake Michigan glaciers, among the most impressive of its kind, formed the boundary between the two lobes for more than 150 miles (*c.* 240 km), the moraine in places being more than 30 m deep.

North America's kettle-moraine, a term fitly given by Chamberlin,[43] is roughly 3000 miles (4800 km) long. It was discovered in Long Island and Massachusetts by W. Upham (1877), in Pennsylvania by H. C. Lewis (1884), in Missouri by G. M. Dawson (1875) and J. E. Todd (1896), and in New Jersey by G. H. Cook and J. C. Smock (1877). Chamberlin found it in Wisconsin (1877) and in his classic paper of 1883[44] traced its course from New York westwards through the states into Wisconsin—in Long Island two belts of terminal moraine traverse the island from end to end, each up to 200 feet (60 m) high and accompanied by an outwash plain to the south. He showed that it was looped about the western and southern ends of the Great Lakes basins. J. B. Woodworth, E. Wigglesworth, M. L. Fuller, A. O. Veatch and F. Leverett, among others, have examined it in later years.

Europe's kame-moraines are traceable for hundreds of miles, as in England[45] and southern Ireland[46] and in the Baltic Ridge of north Germany (see p. 1163). In Scotland, they often occur at the mouths of valleys[47] where the expanded foot glaciers might have been expected to stagnate and decay.

Origin. Kame-moraines are original features: unlike drumlins (see p. 393), they have seldom been credited to postglacial forces acting upon uniform surfaces.[48] Their origin is plainly bound up with that of their most typical feature, the kettle-holes. These have been the subject of much unprofitable discussion and of many hypotheses, most of them untenable and now only of historic interest; such are the references to volcanic craters,[49] rotating icebergs, subsidence or solution of underlying strata[50] or of blocks of soluble rock in the drift[51]; to capricious eddies and currents in seas,[52] in extraglacial lakes[53] or in streams discharging into the sea where an ice-front stood below sea-level,[54] or to the self-raising of lakes without outlet.[55] Equally indefensible is the suggestion, originally made by G. Berendt[56] and E. Geinitz[57] and accepted in part by others,[58] that kettle-holes are "eversion lakes" (*Strudelsölle*[59]), eroded at the bottom of moulins—this conflicts with their size, shallow basin-shape and great abundance, and with the lack of any

signs of this action in modern streams or of arrangement in rows perpendicular to the ice-edge.

Some kettle-holes may be due to pressure from an oscillating ice-edge[60] which caused thrusting over a lower block of ice,[61] to stream scour,[62] or to aeolian redistribution of sand before the lateglacial vegetation obtained a foothold.[63] With much more probability they may be referred to an irregular building up of the drift[64] or incomplete filling of the "creases" of outwash plains.[65] G. D. Hubbard[66] tried to show by experiments with clays in tanks that a pit-forming process is normal with overloaded streams. H. L. Fairchild[67] thought that kettle-holes, created by circumdeposition, were shallow, had smooth curves and walls of finer material, and were situated below the water-table, usually behind bars or in embayments.

But the vast majority of kettle-holes owe their existence to an irregular settling of fluvioglacial deposits over or about melting masses of dead ice left behind at the dissolution, the *Einsturzsölle* and *Aussparsölle* respectively of H. Spethmann,[68] the *Toteiskarst* of C. Troll.[69]. Such masses (promontories, peninsulas and detached island-like bodies represent the successive stages[70]) melted as they lay under the sands and gravels; Troll[71] has suggested that the larger masses should be called "stagnant ice" and only the smaller masses "dead ice". Early geologists arrived at this view from studies in the field,[72] including modern glacier regions in Iceland,[73] or inferred it from the ground-ice of north Siberia[74] (see p. 565). Full confirmation came when kettle-holes in the making were discovered at the margins of Alaskan, Spitsbergen and Icelandic glaciers[75] (in Iceland they are partly vulcano-glacial) and *Dryas* clay was found replacing *mull* locally in Denmark where dead ice prevented this lateglacial layer from forming.[76] Depressions in the lava flows of Ross Archipelago may denote similar slumping following the melting of subjacent ice.[77]

Some kettle-holes may be due to the wasting of stranded blocks of ice on lake-floors,[78] or of those swept from glaciers at the bursting of glacier-lakes.[79] Others may owe their origin to the melting of winter ice on streams issuing from glaciers or springs[80] (see p. 565).

Hugi, Agassiz, Forbes, Tyndall and other students of Alpine glaciers during the advancing phase of the first half of the 19th century (see ch. VI) knew nothing of decaying glaciers from direct observation. Our knowledge of the *död brae,* as Steenstrup[81] named them, comes largely from modern polar investigations. The dead ice may be separate from the glacier or may constitute a stagnant margin to the ice which farther back is still moving, though possibly so slowly that its surface is covered with trees (see p. 912) or with an unbroken carpet of moss.[82] Lenses of dead ice, directly visible or indicated only by a settling or cracking of the ground, by an abundant outflow of cold water or by the weathering of surface debris, are found in almost all modern glacier-regions.[83] In the Alps,[84] they occurred after the retreat of the 19th century and in later years. These bodies, sometimes not quite inert,[85] may reach considerable sizes[86]; instances 800 m long have been recorded from Alaska and 4–6 sq. miles (*c.* 10–15 sq. km) in area from the Antarctic. They may lie on cols[87] or be cut off from the main mass by a divide or high step[88] or by eroding streams[89] (if the ice is thin and crevassed) or by the sea, as on islands in ice-fjords. They may also originate, as S. Finsterwalder[90] suggested and others have observed,[91] by the "tearing off of a glacier tongue" in the rapid recession following an advance. They commonly

originate because of the protection by ablation moraines. Dead ice may also arise from the burial of lake-ice.[92]

Stagnation of a margin still in contact with living ice is known from the Arctic[93] where terminal or inner marginal moraines lay wholly on the inner side of immobile ice.[94] It may be due to deficient supply, as in the Turkestan glacier type,[95] to overloading with debris[96] (see p. 214) on to a topography inimical to flow during a retreat, as in Franz Josef Land[97] where the shoulders of plateaux protrude.

Acephalous ice, or glaciers shrinking at their heads, have been seen in polar lands.[98] In the Antarctic, where instances are provided by the Blue, Garwood, Davis, Ward, Howchin, Miers, Hobbs and Taylor glaciers, the starvation is due to the high rims at the heads of the basins[99]; it harmonises with the extreme cold (see p. 167) and the absence of running water (see p. 450).

How long the isolated stagnant masses survive is almost unknown; it depends upon the depth of the covering debris and the nature of the subglacial drainage. In Alaska, they have been known to last 10–12 years,[100] and in the Pamirs may survive 80 years[101] and in Iceland 200 years.[102] Some stagnant ice, buried under stratified drift, may have persisted in the upper Mississippi valley from the Patrician into the Late-Wisconsin,[103] in Finland for about 1000 years and in Germany, on the evidence in East Prussia, at Meiendorf and in the Oder valley, for some thousands of years.[104]

Isolated mounds may be "perforation deposits"[105] built up in vertical shafts which had no bottom outlets.

Marginal. The origin of kame-moraines is revealed in their composition, structure and kettle-holes, and more especially in their field relationships, a true appreciation of which is frequently more important than minute inspections of exposures. During the four decades after Agassiz's enunciation of the Glacial Theory, the big moraines of the ice-sheets, as distinct from the small moraines of valley glaciers, remained undetected. Recognition came first in North America[106] and later in Britain[107] and north Germany.[108] Kame-moraines are the moraines of the ice-sheet,[109] if the word moraine be used in its broadest sense for an aggregate of drift deposited along an ice-margin; they are unquestionably of morainic nature and mark the edge during dissolution as characteristically as do moraines of mechanical origin.

Melt-waters of subglacial and superglacial origin[110] piled them up against, around and upon the borders of thin ice which in its decay, as was early conjectured,[111] broke up into an irregularly frayed mass, with re-entrant bays and cavities and isolated pieces. Water-worn sand and gravel from englacial sources clogged the hollows and passages in the margin while emerging subglacial streams, which effaced the striations on the glaciated pebbles and transported the muds to greater distances, deposited fans, cones and deltas. If the waters were narrow, the load was laid down as stream-deposits, if broad and tranquil, as lacustrine beds, with laminated silts and varved clays in the deeper parts. Some of the detritus melted out submarginally,[112] to give the *Grundmoränenlandschaft*[113] which was essentially end-moraine in respect to the ice-sheet.[114] Local revivification of the ice produced *stau* disturbances[115] while the final melting caused the materials to slump, with a resultant folding and faulting and a dipping of the stratification into the hollows.[116] The irregular lumps of till, up to 25 cm long, which are more or less rounded or

oval in shape and occur at the rim of a kettle-hole in an outwash plain, were formerly associated with the ice which was responsible for the hole.[117]

The theory that kame-moraines arise from stagnation of an ice-margin has withstood the objections that have been raised[118] by those who believe the materials are too thick and the kettle-holes too plentiful; that stagnant ice would not deliver the requisite volume of water; or that the older drifts, often regarded as products of stagnation (see p. 1150), have no holes.

Kame-moraines, therefore, were amassed along a frayed and immobile margin by countless distributaries unloading their burden in crevasses and embayments and between detached masses of ice.

3. *Osar*

Topography. Osar (Swed. *Ås*, sing.; *Åsar*, pl.)—the suggestion[119] that the anglicised form should be *ose*, plural *oses*, has not been welcomed—are tortuous or gently sinuous ridges ("serpent kames"[120]), sometimes hundreds of kilometres long.[121] Their appearance is often remarkably artificial, resembling railway embankments (pl. XVI, p. 432). In consequence, they were early attributed in North America to aboriginal races and denoted Indian ridges.[122] Their crest is smooth or broadly hummocky and typically so narrow that it can just accommodate a path or road (Ger. *Wallberge*). The cross-section,[123] like that of a goat's back, is roughly an isosceles triangle; the sides are steep or abrupt and sometimes, e.g. in the Central Plain of Ireland, asymmetrical.[124] They rise to over 45 m above the adjacent country; in New Jersey,[125] they average 45 m broad and 4·5–7·5 m high and in Finland[126] 20–30 m high, exceptionally, as in the Kangasalås 80 m and in the Sareks region 85 m.[127] The Uppsala Ås[128] reaches an absolute height of 140 m, of which later clays conceal more than 100 m.

Osar are often composite and split into two or more parallel ridges or "double eskers" enclosing kettle-holes (Swed. *Åsgroper*) and irregular basins. Cross ridges frequently interlace them into "reticulated ridges" or "os-nets". They may widen out into plains or "plateau-osar". The summit line may be uniform and even but more often is wavy along the alternate sags and swells, the "esker knobs" or "nodes"[129] (*Åscentra*[130]). The sags in such "beaded osar" may be so deep that the os breaks up into a chain of separate mounds, either aligned or laterally displaced (De Geer's *Kastingar*). The knobs frequently coincide with the bigger changes in the course.[131]

Osar are often bounded by bedded sands and clays and by narrow lateral moats or ditches, sometimes occupied by peat flats or "esker ponds".[132] These "os troughs"[133] (Swed. *Åsgrafvar*; Ger. *Osgrüben*), first noticed in Sweden by W. Hisinger,[134] skirt the os on either flank or oscillate from side to side[135] or divide individual ridges from each other. An intermediate ridge of till or of sand and gravel may subdivide them into two channels.

Structure. Osar consist of water-worn sands, which preponderate in the wider expanses and the submarine forms, and fairly coarse gravels (Swed. *Åsgrus*)—hence the term *Rullstensåsar*—which are often "openwork" gravels[136] (diakene), the empty spaces, with little sand to fill the voids, suggesting rapid accumulation[137] or gravitational deposition.[138] Striations are rare except on an occasional large boulder, though faceted and flattened boulders are numerous and lie across the os[139] and more or less horizontally[140] in a way that hints at washing rather than rolling.[141] A few osar are made of

till, as in Woodworth's "false eskers",[142] or are coated with or enclose large masses of it.[143] Others have cores of solid rock.

The materials[144] differ little in composition from the local boulder-clay and contain a large local contribution, though the percentage of far-travelled rocks is sometimes higher. They are rudely and irregularly stratified, current-bedded and ripple-marked,[145] and their coarse and fine constituents including laminated clays alternate rapidly. A periclinal or anticlinal structure ("over-cast bedding"[146]) is common though by no means a constant feature.[147] Like the lateral faults which hade outwards, it is often made by settling as the ridges adjusted themselves to the vanishing walls of ice[148]—artificial excavations produce identical longitudinal structures[149]—though some of it is due to undisturbed deposition at angles oblique to the bedding in the centre of the mass. The coarseness varies much along the line of the os; the proximal end is coarser, the distal end finer.[150] The direction of the foreset beds and of the decrease in grain size, together with the orientation of the material, indicates the direction of the stream which produced them.

Distribution. Osar are usually short, stretching for only a fraction of a mile as in Connecticut and north Germany. Yet they sometimes and farther north have lengths, as in Maine,[151] of 160 or 240 km, and in Scandinavia[152] of 300–400 km or even 450 km in the case of the Söderala-Uppsala Ås. Larger osar are often disposed like the tributaries or distributaries of a river-system,[153] the former (Swed. *Biåsar*) being inferior in development and length to the main osar (Swed. *Hufvudåsar*) which they meet acutely, though the tributaries often run for considerable distances parallel with the main osar before joining up.

Osar usually run from higher to lower levels, especially in valleys where they lie in the axis or, more normally, on one side.[154] They may cross or recross a valley or proceed down one side and up the other[155] in an "uphill and down-dale" manner. The supramarine forms may even traverse hills and valleys,[156] usually by low passes,[157] beyond which they are continued as river-eroded channels.[158] Nevertheless, their topographical independence is limited.[159] This is readily explained by their general parallelism with the local ice-flow ("radial moraines"[160]) as fixed by drumlins, boulder-trains, erratics and striae and observed at an early date[161]: if the ice moved from different directions the osar may cross[162] (see p. 354). Naturally they must traverse the intervening ridges if the valleys are oblique or perpendicular to the flow, though they are often broken on the watersheds. Occasionally, they depart from this direction as in the case of the *Biåsar*: they may even run parallel with the ice-edge and moraines[163]—these are the "cross osar" (Swed. *Tvåråsar*; Ger. *Querosar*) of Sweden[164] or "marginal osar"[165] (Ger. *Gerollenmoräne*[166]; *Kiesmoräne*[167]). *Langåsar* on reaching the ice-margin may turn parallell with it[168] or may be monoclinally stratified towards the lee.[169] Osar may indeed resemble end-moraines. Their sinuousity (sinuous moraines about modern glaciers are not unknown[170]) and their relation to the ice-flow are, however, diagnostic.[171]

Osar may end abrupty or gradually, or as "feeding eskers"[172] ("proximal osar"[173]) may end in kame-moraines or in wider tracts,[174] the "kame-plain" or "osar plain"[175] (*Rollsteinfelder*[176]). They may also terminate in sub-aqueous "esker deltas" or subaerial "esker-fans".[177] Osar have also been found at the outlets of glacier-lakes.[178]

The more or less special conditions seemingly necessary for their birth

cause osar to be relatively rare. Their fullest European development is in Scandinavia[179] (fig. 74). Here they mainly occur east of the Pleistocene iceshed, particularly in central and southern Sweden—they sometimes appear on the adjacent sea-floor.[180] In the higher parts, they are strictly

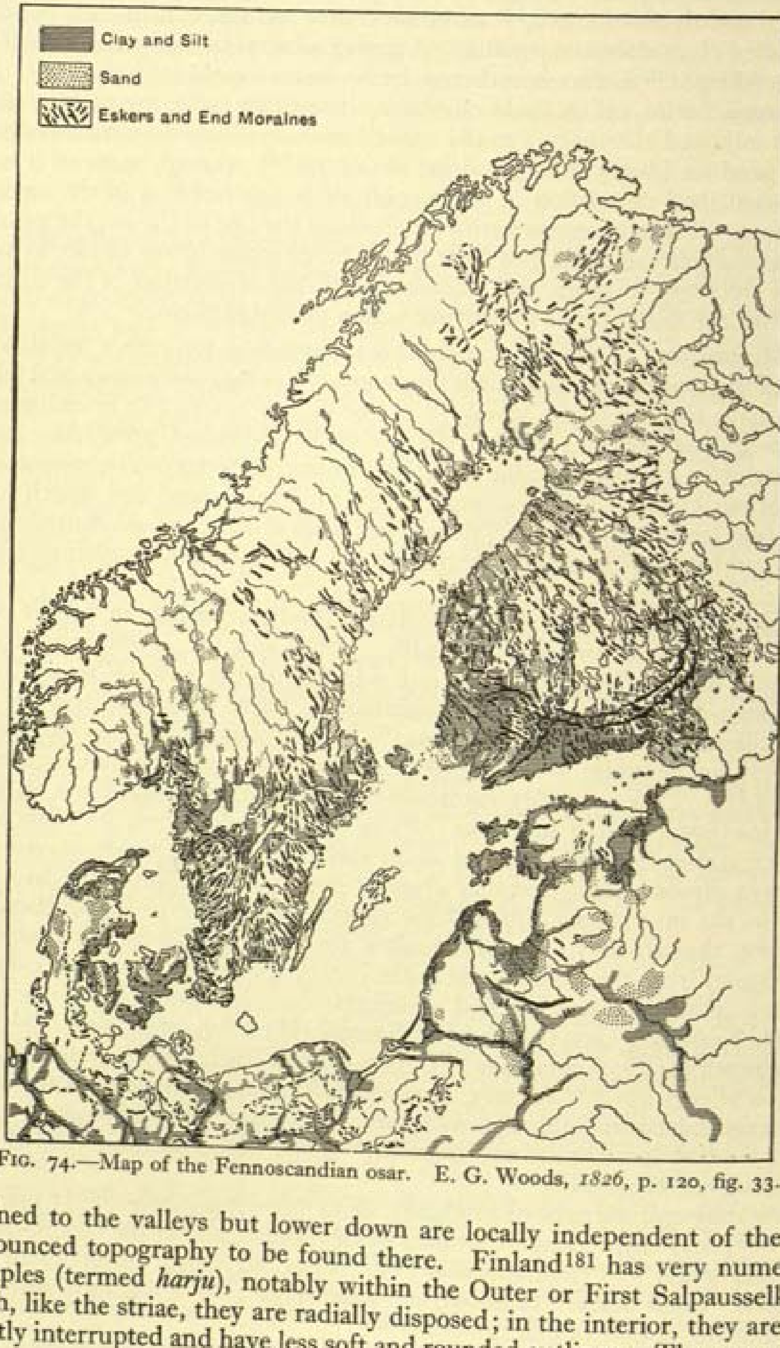


FIG. 74.—Map of the Fennoscandian osar. E. G. Woods, *1826*, p. 120, fig. 33.

confined to the valleys but lower down are locally independent of the less pronounced topography to be found there. Finland[181] has very numerous examples (termed *harju*), notably within the Outer or First Salpausselkä to which, like the striae, they are radially disposed; in the interior, they are frequently interrupted and have less soft and rounded outlines. They are much

rarer in Norway[182] but have been discovered in Jaeren and the west; they include the *egger* (edges) or sharp crested ridges which follow the axes of the valleys.[183]

The Kola Peninsula is apparently without osar[184] though they are found between Lake Ladoga and the White Sea and east of the Baltic[185]: here they are shorter and less regular than in Sweden and lie within or north-west of a drumlin zone. They occur too on the islands and in the eastern part of Denmark[186] where they often lie in the tunnel valleys or in line with these, and in many parts of north Germany[187] (Ger. *Wallberge*) where attention was directed to them at a much later date.[188] Examples are not unknown in the terrain of the Alpine glaciation[189] and a number have been described from the British Isles: some are among the diverse accumulations J. W. Gregory[190] listed. Central Ireland has numerous examples (see fig. 250, p. 1210).

North American osar are best developed in Maine[191] (fig. 75) but are known from other states[192] (Wisconsin, Minnesota, Michigan, Illinois, New York and New Jersey). To them belong the *ispatinows* of north Saskatchewan[193] and the eskers of "Lake Agassiz" and other parts of Canada[194] including those east of Great Slave and Bear lakes, west of Hudson Bay and south and east of James Bay, in central Quebec, and in Labrador. As in Finland, the long sinuous ridges stand out sharply from the multitude of lakes and monotonously uniform drift or solid ridges. In the region about Great Slave Lake the presence and the size of the eskers appear to depend upon the existence and amount of local ground-moraine.

Origin. The various hypotheses propounded for the osar reflect the changes of thought that have succeeded each other about glacial events. Like most of these, they are now in the main only of historic interest. This can safely be said of the views that osar were laid down in the lee of obstacles by a Northern Flood,[195] by marine and tidal currents,[196] or by floe-ice or icebergs floating in a glacial sea,[197] the Esker Sea[198]: Kinahan[199] classified them as "fringing", "barrier" and "shoal" eskers. Only less erroneous are the hypotheses which derive osar from a melting out of median or inner moraines[200] or from push moraines[201] or regard them, on the "strandwall theory", as the limits of a glacial sea. Equally indefensible is the view that they were postglacially denuded out of once-continuous sheets of drift[202] ("pseudo-osar"[203]). In Denmark they were associated with a goblin with a leaky sandbag.[204]

In statu nascendi. The exclusion of all these hypotheses still leaves much room for uncertainty which research on existing glaciers has done little to dispel: modern accumulations resembling osar in appearance and structure are singularly few. The investigator is baffled at every turn. The Antarctic ice-sheet, perhaps the nearest analogue of the Pleistocene ice, protrudes almost everywhere beyond the land into the sea so that any accumulations there may be are inaccessible and severe cold reduces drainage to a minimum. Search along the ice-fronts in Greenland, though rewarded by an occasional discovery of fluvioglacial material,[205] has on the whole been fruitless[206]; the streams are inadequate, median tunnels are absent (drainage is usually lateral), and the rocks are soft. It is true that a winding ridge of sand and gravel, 6 m high and about 80 km long, running parallel with the ice-flow, has been noticed[207] while a similar ridge, 4–5 m high and 10–30 km long, was seen in east Greenland.[208] In King's Bay, Spitsbergen,[209] an os-like ridge was left

when a sheet of gravel, deposited in a glacier-lake, was eroded.[210] The flat and uncrevassed part of Norway's Lodalsbrae had a median moraine which was over 2 km long and over certain stretches lay in a valley-like depression that might have produced an os-like feature at melting.[211] The Böverbre of Jotunheim had a ridge in process of formation which was 36·6 m long and

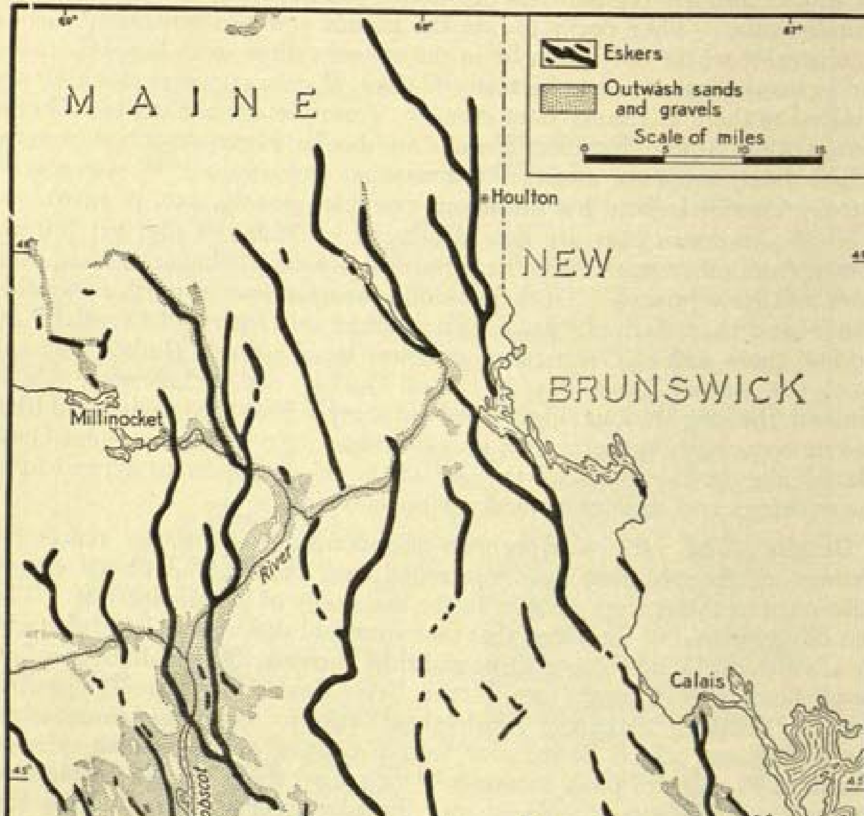


FIG. 75.—The osar and outwash deposits of south-east Maine, U.S.A. W. W. Atwood, *83*, p. 91, fig. 44.

3·6 m high.[212] There was a tiny os, the product of a subglacial stream, at the snout of the Syter Glacier, south Lapland.[213] A sand and gravel ridge, 100 m long and 3–4 m high, was formed in an englacial channel at the lower end of the Lower Aar Glacier.[214] Recent os-like features have been observed north of the Vatnajökull[215] and in front of the Morsárjökull (Ives & King,

1955) in Iceland and in the Wind River Mountains of Wyoming.[216] In Alaska[217] they comprise an os with small tributaries in front of a glacier of Yakutat Bay, a doubtful subglacial os in front of the Woodworth Glacier, and accumulations built up by superglacial gravels sliding through holes in the roof of broad subglacial tunnels. A small ridge was connected with a short glacier on Mount St. Helena, Washington[218] (pl. XVIIA, p. 464).

Even where circumstances seem to encourage their formation, osar are later destroyed through the agency of the very stream that created them. Thus the vast gravel spread which covers hundreds of square miles in front of the Malaspina Glacier had only one feature resembling an os,[219] and a ridge, 15 m high and 90 m long, noticed by C. W. Paijkull[220] in Iceland, had disappeared when the locality was visited a few years later.[221]

Modern Alpine glaciers, notwithstanding their median drainage, have no osar: high gradient and inadequate detritus with which to clog the outlets of the channels are apparently inimical.

Modern glaciers, therefore, have contributed little to elucidate the os; no process or form has been discovered that throws any definite light upon the problem. The glacialist is driven back upon a careful study of Pleistocene accumulations and a severe testing of any working hypothesis it may suggest. Hence the contradictory views which are still prevalent. Nevertheless, the testimony of modern ice is not quite negative; the belief is confirmed that an os is built up by a glacier stream flowing along its length. This harmonises with the linear extent, the rolled gravel, and the thorough elimination of the fine silty particles; the tortuous courses with loops and knobs and ramifications reminiscent of streams; the traces of alternate erosion and deposition along the os chain; the frequent parallelism of the boulders with the axis of the os[222]; the current-bedding directed towards the distal end[223] (De Geer's *uppström* bedding); and the arrangement of osar within the Salpausselkä, related to longitudinal crevasses or the domed surface of ice-lobes.[224] Modern ice too shows that the os, whether sub-, en- or superglacial, was born like the kame-moraine and the drumlin in the lower layers of the ice and gathered its material from englacial sources.[225]

Supporters of the subglacial or superglacial hypotheses presently to be mentioned have consistently considered an inert margin essential for os formation[226]: the finely developed osar in Maine and Sweden may be due to stagnation arising from the critical relation between the general slope of the land and the minimum gradient for effective ice-flow during the closing phases of dissolution.[227] The following seem inconsistent with any appreciable forward movement: the irregular outline of the ice as depicted by marginal accumulations; the want of evidence of glacial thrust; the wavy crest and the winding of the ridges; the splitting of the osar which involves the presence of masses of dead ice or the erosion of tunnel walls and the isolation of pillars of ice; the inconceivability of open channels being maintained athwart the flow—tunnels driven in modern glaciers across the down-valley flow close in fairly rapidly by a "bulging" from the upstream side[228]; and the association of the marginal ditches with a slumping of the gravels when the supporting ice-walls melted away.[229]

Yet some forward motion, apart from the features inherited from an earlier active stage, may have taken place in some instances; the osar are parallel with the flow which guided the courses of the glacier-streams,[230] either superglacially by yielding a frontal slope or subglacially by closing any channel

oblique to the flow; stream eroded channels lie beneath them[231] (see p. 428) and till rests upon them[232]; drag is sometimes seen along their margins[233]; and some osar have been subject to torsion.[234]

Superglacial hypothesis. A view which was simultaneously enunciated in North America and Europe and has commanded substantial approval[235] and powerful advocates in Stone[236] and Crosby[237] regards the os as constructed superglacially from debris in surface canyons licked out of immobile and uncrevassed inframarginal ice. Basal melting and subglacial streams lowered the embryonic os to the ground without loss of its distinctive features, though faulting and disturbed stratification attended the action. Slumping induced the irregular crest. The superglacial streams were limited in length only by the breadth of the zone of ablation in which englacial debris became superglacial by melting. In the few cases where crevassing drained off the superglacial streams, the interstream surface was lowered more rapidly than the old gravel-filled channel and the esker material gave rise by sliding to the double esker.[238] Reticulated eskers or plateau osar were built up in rapidly enlarging lakes, open to the sky, or in several parallel but interconnecting channels, with delta-like branching, if the surface was nearly level and the water flowed more slowly.[239] Esker knobs denote confluences with hanging tributaries[240] and cross osar cross crevasses.[241]

The following are cited in support[242]: few boulders are to be found on osar—ice floated these in open channels; osar broaden towards their termination as the river channels widened; and the minor or larger reaches of the meanders are rarely related to the local topography. The rival (subglacial) hypothesis (see below), it is claimed, does not well explain the characteristic cross-section, the occurrence of cross osar and the variation of shape longitudinally. It also encounters the difficulty that osar, particularly in the reticulated type, are too wide and are not specifically developed where crevasses were likely. Moreover, subglacial streams would inevitably be forced up into the ice as the roof melted to conform with the base level of either a lake-surface or a detrital delta (see below).

A. W. Giles[243] has very fully set out the grave objections to this superglacial view. Thus surface streams on glaciers, as often observed,[244] are usually free from debris since their sides are too steep, their sides and floor are too smooth, and their flow is too swift to permit debris to lodge, though water-worn pebbles and detritus have occasionally been seen[245]; many osar are longer than any probable stretch of uncrevassed ice; glacier-drainage at present is subglacial; osar and os troughs lie in depressions in till or solid rock (see below); and the simple linear outline would scarcely be retained on collapse.

N. H. Winchell,[246] in a modification which has found not a little favour,[247] thought osar filled radial crevasses in dead or live ice (in crevasses parallel to the ice-edge cross osar arose[248]). The crevasses were opened by rock-obstructions (some osar occur in the lee of such features[249]) or by differential movement along the plane of contact of two lobes[250] that alternately widened and narrowed the os ridge. The mammalian bones occasionally discovered in osar[251] have been regarded as confirmation.

Against this modification may be urged the objections already raised against the major hypothesis, including the great length of the osar, as well as others peculiar to itself, such as the winding courses of the osar and their occurrence where crevasses were unlikely.

R. F. Flint's crevasse infillings[252] were accumulated on or near the edge of a decaying ice-field, as noticed earlier in New Jersey[253] and on modern ice in Spitsbergen.[254] Their margins are scalloped where promontories of ice projected into ponded water and the material slumped to the angle of rest. They may embrace the "reticulated eskers" and "lone kames"[255] and are closely related to recessional moraines and pitted plains but lie beyond the former and are incorporated with the latter. In comparison with osar, they are usually small and short (say, 400 m or less) and low, usually 3·5–4·5 m only. Characteristically, they do not wind nor do they possess till-coats, knobs or a tributary plan. They run continuously and regardless of the flow-direction and lie downstream from ordinary osar in more fissured ice. Many German osar may have originated in this way.[256] Marine crevasse infillings in a stagnant ice-sheet have been described from Quebec.[257]

Subglacial hypothesis. A widely accepted hypothesis[258] regards the osar as aggraded by subglacial streams which flowed under considerable hydrostatic pressure and coursed in ice-walled, tube-like tunnels, with arched roof and wide base, even in submarine areas.[259] They portray the pattern of the subglacial drainage, especially in the final inactive state of the ice: their streams were fully loaded unlike those of the *Rinnentäler* which were under-loaded and eroding.

Drainage under the centre of the ice-sheet, though present on account of the earth's heat[260] and possibly in the form of a "pressed melt",[261] was in all probability insignificant save during the last stage[262]: the perfect and continuous contact of ice and bed (as the universal striae indicate) and the absence of basal crevasses and moulins seem to point in this direction. Such signs of subglacial drainage as giant-kettles (see p. 238), stream-eroded channels in till and rock (see below), *Rinnenseen* (see p. 241), and sands and gravels in boulder-clays are, from modern analogies, referred to the marginal zone of melting[263] in which the diffused subglacial waters of the inner zone became concentrated into definite channels and were led by marginal crevasses into the subglacial drainage system.

These subglacial streams, derived from superglacial sources, springs and frictional heat, followed the inclination of the ground where the constraining ice permitted, or strove towards the regions of lower pressure at the ice-edge—a backward drainage may sometimes have existed[264] (see p. 469). They enlarged their channels by mechanical erosion and by melting their walls, and worked their way upwards by heaping debris on their beds and at their mouths, as observed on Alaskan glaciers[265] (see p. 436). In this way, they became englacial or even superglacial over part of their course.[266]

Osar were born in this marginal melt-zone which had a width possibly of a hundred to several hundred miles[267]—in modern Greenland it occupies about one-sixth of the ice-sheet and in the Scandinavian ice-sheet was probably 180–200 km broad—though the tunnels at any one time may have been only a few, say, 2–3 miles (3–5 km) long.[268] In other words, osar were not built up simultaneously from end to end; the proximal end was much younger than the distal end.[269] Glacier-streams in North America may have been shorter than in Scandinavia because the ice receded more slowly and both topography and crevassing of the thin ice were less pronounced,[270] though in both Europe and North America a greater continuity and higher proportion of contained erratics suggest that the stagnant ice zone widened northwards as thinning progressed.[271]

This subglacial origin has much supporting evidence: the material has been carried uphill[272] and is compact and local[273] (it may attain 98% of the bulk), especially if the os is small or only moderately long; erratics and boulder-clay occur upon the summits and sides[274] (this is not crucial[275]); the bedding is arched; and the hydrostatic pressure, denoted by the size of the blocks, was great. Moreover, osar are parallel with the striae and ice-flow, and the surfaces beneath or associated with them are water-worn.[276] They repose in river-like channels in boulder-clay or solid rock,[277] usually many times broader than themselves and occasionally pot-holed—these channels may be much more frequent than observation suggests since exposures are seldom deep enough to reveal them; erosion is sometimes seen in the line of the osar between sections,[278] as on the distal sides of divides (see p. 421) and is demanded by the os troughs[279]; and river channels, connected with osar and up to 30 m deep (they number over 300 in Massachusetts[280]), run parallel with the ice-flow on the crests and flanks of drumlins.[281] Osar are associated with *Rinnenseen*, either along their length or by their side[282]—those within the tunnel valleys may be erosion residuals or "false osar"[283]—occur far below the level of lakes[284] or of the sea, even where the ice was standing in water 200 m deep,[285] and are related to moraines and kames.[286] The Woodworth os of Alaska (see p. 425) shows that this origin cannot be neglected. The boulder-clay sometimes seen may have been due to pressure from the sides or the base.[287]

The os may have originated at a moulin,[288] especially downstream from an obstruction which initiated crevasses (see above). Spurs may mark the sites of alcoves in the ice[289] or small leakages from the main stream,[290] while double eskers may have arisen in a main channel kept clear by a swiftly flowing river[291] or in a widening of the tunnel in which the arch sagged and divided the stream into two branches.[292]

Breaks in the os or their replacement by large pebbles[293] are attributed to the stream's more rapid flow downhill near heights of land or in narrow places. The os centre on the other hand is linked with superglacial material falling into holes in the tunnel, the enlargement of its sides and roof,[294] the stream's greater velocity in restricted passages (Woodworth[295] emphasised the channel's influence upon the height of the os), or with deposition in wider and quieter reaches and the irregular slipping that resulted when the ice-walls were removed.

Others think an englacial stream, heavily charged with sand and gravel, originated the os.[296] It was tapped by proximal crevassing and the debris settled about the melting ice-core with faulting of the material and arching of the bedding. Asymmetry of outline was due to aspect. When a transverse crevasse was opened, the coarser material was deposited first as the os centre, the finer material being carried forward. On this view, the tributary os arose from crevassing on either a parallel or a transverse course; the os trough was eroded when crevasses penetrating to the base tapped the main stream; and the os ridge after subsidence came to rest on the side or across the trough.

Rinnenseen and osar, with their right angled pattern, mark, it has been thought, the net of the final or "fossil" crevasses which "granite-tectonically" were directed perpendicularly to the longitudinal crevasses associated with push moraines and *Urstromtäler*.[297] The *Aufpressungsosar* of north Germany may have been squeezed into radial crevasses from below by the weight of ice.[298]

Frontal delta theory. In a theory almost simultaneously put forward by O. H. Hershey[299] in Illinois, by De Geer[300] for the osar of Uppsala, Stockholm and Dalsand, and by E. v. Toll,[301] the essential condition for the origin of osar was the sea or an extraglacial lake standing in front of a retiring glacier. The overloaded subglacial stream deposited its debris at its mouth, as at the edge of the Malaspina Glacier to-day,[302] or in "calving bays" possibly 1 km or more long.[303] At the recession, the "esker-deltas" arranged themselves in successive series as a more or less unbroken ridge, the "radial os", whose windings conformed with those of the subglacial river. The result was a continuous ridge like a long "squeeze" of tooth paste. If the ice-border remained stationary for any length of time, the deltas grew up side by side as a marginal os, especially when the channel mouths were clogged and laterally displaced.

This theory had already been anticipated: W. Upham,[304] for instance, suggested that osar were progressively deposited at the mouths of glacier-streams as the ice withdrew. Others recognised that osar were particularly good where streams discharged into the sea in marginal bays.[305] De Geer, however, completed the theory by relating the knobs to excessive melting in summer, the sags of finer sands and clays to diminished flow in winter, and by bringing out the thickening of the varves about the os centre (the significance of this had been earlier anticipated[306]), as shown by isopachytes (disposed in irregular and semicircular lobes with interlobes) in the floor deposits.[307] A. C. Trowbridge[308] thought a gap occurred if the re-entrant disappeared, and the os broke up into an intricate maze of ridges where the front had several re-entrants or was very irregular. De Geer correlated the knobs with annual moraines (see p. 1151), though H. W:son Åhlmann[309] equated them with the accumulations of autumn, winter and spring, when the ice was stationary, and linked the sags with the summer retreat.

The delta theory has had wide approval[310] and some experimental support[311]: it is consistent with the evidence of ponded waters, the well-defined cliff-like edge of ice ending in standing water, the rapid wastage that occurs under these conditions, and the preservation of the os-segments when they are abandoned.[312] It has, however, been objected to[313] because the hydraulic pressure was insufficient, the materials were built out as in a normal delta and not upwards from a subglacial stream, and on other grounds. The crucial test is whether the crest does coincide with the level of the standing water. Osar very often agree closely in height with other glacial deposits, e.g. moraines and deltas, and are related to and controlled by the altitude of the lateglacial sea in many parts of north Fennoscandia[314] or of extraglacial lakes,[315] as in Finland and Estonia. Elsewhere,[316] as in supramarine Fennoscandia, north Germany, Quebec and Maine, they were built up above the level of large, continuous bodies of water. Such supramarine ridges are usually steeper than subaquatic ones[317] and in Sweden are wider, have a more irregular top and less rounded or sorted material,[318] and change their shape if they cross the marine limit.[319]

Conclusion. A critical discussion of the rival hypotheses shows that the problem of the os awaits its final solution. The general characteristics are readily ascertained; interpretation alone is difficult. Most probably, more than one theory is essential to a complete explanation and it is inadvisable to premise that all osar arose in one and the same way. As in other cases, care has to be taken not to overdraw conclusions. Osar may apparently have

formed as on the delta theory in the sea or ponded water or at the mouths of or within subglacial tunnels[320] (especially if the streams had insufficient velocity to carry their full loads uphill from beneath the ice to the higher parts of deltas against the back pressure of the standing water) or by superglacial streams without such control.[321]

Though the features are polygenetic, they represent the waning stages of glaciation (the Esker Period, to use E. Durocher's term[322] in another sense) after the drumlin period and were accumulated in the marginal zone by glacier-streams fed by melting. These, possibly in part superglacial and englacial but also subglacial, deposited their burden within and at the mouths of tunnels. Whether the os was superaquatic or subaquatic depended upon the incident of the proximity of standing water. Each os must be separately diagnosed, paying due respect to its external form and internal structure and to its relation to the waning conditions in its own area.

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206. **132** *Ibid.* 213. **133** F. Leverett, *A. J. S.* 142, 1891, 209. **134** *764*. **135** P. G. Krause, *J. LA.* f. 1911, I, 76. **136** W. M. Davis, *Boston S. P.* 25, 1892, 489; J. Elbert, *JB. Gg. G. Greifsw.* 8, 1904, 38. **137** W. M. Davis, *op. cit.* 476. **138** H. Wadell, *G. Ann.* 18, 1936, 74. **139** G. Lundquist, *S. G. U.* C. 497, 1948, 11 (fig.). **140** *1644*, 463. **141** J. W. Gregory, *T. G. S. Glasg.* 14, 1913, 208. **142** *Op. cit.* **143** *1644*, 469. **144** *37*, 288; *467*; *1644*, 444; *1646*, 40; W. M. Davis, *op. cit.* 477; A. Hellaakoski, *F.* 52, 1930, 1; W. G. Simon, *Z. D. G. G.* 89, 1937, 76; J. M. & H. B. Trefetlen, *A. J. S.* 242, 1944, 521. **145** *970*, 98; *1598*, 819. **146** *1598*, 819. **147** R. F. Flint, *op. cit.* 411. **148** *37*, 133, 286. **149** R. F. Flint, *op. cit.* **150** *385*, 382; *1308*, 68; *1644*, 430, 465; *1821*, 95; G. De Geer, *S. G. U.* C. 173, 1897; J. Elbert, *op. cit.* 39; P. G. 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A. Kame-terraces, Madawaska Valley, Ste Rose, Quebec
[R. J. Lougee]

B. Kame-moraine, Carstairs, Lanarkshire, Scotland
[Geol. Surv. Gt. Britain : Crown copyright]

Ose ridge crossing River Dalaloen, Sweden [A/B Aeronautic, Stockholm]

(*facing p.* 433)

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Chapter XXII

VALLEY TRAINS AND OUTWASH PLAINS

Topography. Outwash fans, sheets or aprons[1] ("frontal aprons"[2]) constitute broad plains whose surfaces are generally smooth and even though sometimes gently undulating: slight irregularities are caused by inliers of ground-moraine, such as drumlins[3] or the *Bakke-Öer* of Denmark (see p. 944). They are indented by kettle-holes, as in the pitted plains (see p. 440), or wrinkled by abandoned stream "creases",[4] "furrows"[5] or "channels", usually narrow and reticulated and closely spaced. The furrows may cut across the kettle-holes which were clearly then still occupied by ice.[6]

The outwash in valleys generally extends from side to side and inclines slightly downstream, filling up any inequalities in the floor and burying the solid rock so that this is rarely seen except in cases of glacial epigenesis (see ch. XXV). The average slope in north Germany[7] is 1:1000 or falls from 1:300 or 1:500 at the head to 1:1000 at the distal end. The gradients in north Switzerland[8] are about the same (they are, below Aarau, 2:1100, above Basel, 1·2:1000, and about Bodensee, 1:330–1:200) and of the same order as in present-day Alaska.[9] The Mashpee pitted plain (see p. 441) slopes at 25 ft/mile (1:210) close to its apex and about 12 ft/mile (1:440) in its lower half. In a linear succession of moraines and outwash plains, the inter-morainic stretches may have a modified gradient or pass into horizontal lake-floors.

Structure. The materials of outwash plains are noted for their heterogeneous size and grain and the constant interchange between clay, gravel and sand (*Mo* sand of some Swedish geologists). The gravels are well rounded and often tumultuous and usually coarser than in a normal river. In places they are poorly sorted and pass both laterally and vertically into sandy and gravelly boulder-clay or into unsorted till. They contain boulders and cobbles at the iceward end and become finer and finer as J. G. Forchhammer[10] noted in 1847 for Schleswig-Holstein. Meanwhile the fan itself thins progressively towards its outer margin and the gradient becomes parabolically less; the Munich Plain[11] falls from 12 to 13% near the ice-edge through 7·7% and 4·4% to 2·2%, and the outwash of north Schleswig[12] from 2–2·5% to 1·4% and finally to below 1%. The materials decrease in grain size towards the distal edge and are arranged parallel to the surface but vertically are variable in texture.

In Denmark, the materials are of two sizes: the sand grains which have been carried in suspension and the small flat stones which were rolled along by the bottom current.[13]

The gravels, which have a higher percentage of foreign rocks than the local boulder-clay, are sometimes consolidated if the underground drainage is checked (as was early noticed in East Anglia[14]) by the lime which is present, giving rise to "calcrete"[15]; ferruginous or siliceous cements make "ferricrete" or "silicrete" respectively.

Outwash of modern glaciers. Outwash plains are rare about the present-day ice-sheets, for in Greenland the ice is largely contained behind mountain ramparts and the Antarctic has no periglacial area. They are, however, in process of formation in places in Greenland[16] and notably in Iceland,[17] e.g. in front of the Vatnajökull, "the glacier that gives water". Here, in the classic area of modern outwash plains, described by O. Torell, K. Keilhack and N. V. Ussing, modern or Pleistocene moraines are rarely seen though then the outwash spreads from the moraine. Their place is taken by bare and wide sand plains, named *Sandur*, pl. *Sandar*[18] (this spelling of the official Danish maps replaces the obsolete form *Sandr* which Keilhack[19] earlier introduced into glacial literature), the largest of which, the *Sprengisandur*,[20] situated east of the Hofsjökull, covers 725 sq. miles (*c.* 1880 sq. km).

The *Sandar*—the word signifies sand—extend, with a slope of 4·5 : 1000,[21] from the long reaches of shingle which cover the glacial terminations, the amount diminishing away from the glacier—the *Sandur* before the Hoffellsjökull falls from 1:60 to 1:110 and 1:140.[22] Water wells up almost vertically in front of the ice proving that the *Sandur* plain is considerably higher than the base of the glacier—in the case of Hoffellsjökull it was 15 m higher than the glacier 10–20 m away.[23]

The *Sandar* are traversed by turbulent and rapidly shifting streams, the "white rivers" or *Hvitáar* (D. Hummel[24] named their deposits *Hvitásand* and others have used the term *Hvitáglacial*[25]). They anastomose as "braided streams" which swing down the valley in an intricate lace-work of interosculating channels (pl. XVIIB, p. 464); together with washouts occasioned by the breaking out of glacier-lakes and oscillations of the ice itself, they present difficult problems to the civil engineer[26]; pastures and farms are swept away; structures are destroyed; and channels, spanned with bridges, are left dry. On days of high insolation and in periods of diurnal high water, when the snowfall of half a year may be discharged during a few spring weeks, the plains, noticeably near the ice-edge, are submerged beneath swollen floods or a myraid of tortuous streams and often form quicksands—similar quicksands may explain the large number of proboscidean bones sometimes found in Pleistocene outwash.[27] Farther downstream, the waters collect into definite channels or strands or mingle with those which, after percolating through the sands and gravels, are thrown out by the increasingly fine material as springs.[28] The overburdened streams, which carry ice-rafted blocks[29] of considerable size and coarser material than those not fed from melting ice,[30] choke their channels, build natural levées, and aggrade the plains. These are again dissected when the load gets less and degradation replaces aggradation. The whole process is so irregular and untrammelled that it baffles all efforts to make a geochronology on the lines of the varves (see p. 1155), though such attempts have occasionally been made.[31] In winter, ice forms along the stream courses and the entire plain may freeze over.[32] Fine clays are laid down in temporary lakes held up by bodies of stagnant ice or masses of detritus,[33] though the electrically poor waters cause the sandy facies to be more important than in non-glacial streams.[34]

Similar plains have been observed in other Arctic lands[35] (Greenlanders term them *Sioraq*[36]), including Alaska where they grade downstream into glacio-marine deposits (see p. 479). They are less common, however, since flat stretches are few and the sea is near.

The *Sandar* are barren[37]: humus is wanting; cold waters percolate through them; floods constantly pour over them; and icy winds blow over them.

Mode of origin. Outwash plains, É. de Beaumont's *terrains du transport ancien*, L. A. Neckar's *alluvium ancienne* and *alluvium glaciare*,[38] G. H. Kinahan's "glacialoid drift",[39] have been variously interpreted as deposits of lakes pent up behind ice-dams,[40] a sandy facies of ground-moraine,[41] or a product of the washing of boulder-clay[42] (*moraine profonde remanié*[43]; *Diluvium remanié*[44]), remnants of which are to be found as islands of boulder-clay, balls of clay, or big boulders. They have also been derived from the englacial material of the melting ice.[45] However, O. Torell[46] in 1857 and K. Keilhack[47] in 1884 showed the resemblance of the north German outwash to the Icelandic *Sandar*, though in Iceland volcanic heat and the *Jökullhlaup* (see p. 65), which give rise to vulcanoglacial sediment,[48] are additional complications. The *Sandar* south of the Vatnajökull, with an annual precipitation of 140–180 cm, are less comparable with those of north Germany than are those on the north with an annual precipitation of *c.* 60 cm.[49]

Melt-water streams, which are laden with the grindings of the glacial mill and pour over or breach the moraines or issue from cones at the orifices of subglacial tunnels,[50] are overloaded despite their size since there is no vegetation, except in patches or bolsters, and the ground is frozen. They drop their debris in the order of its coarseness so that the recession of the ice and its streams leads to coarser sediments being buried under later and finer ones. The sheets were probably built up fairly rapidly,[51] particularly if toll was being levied on higher and earlier gravels. Their gradients were such as the size of the streams and the loads demanded.

Striated pebbles are seldom found in the streams of modern glaciers, especially if these are retiring.[52] Constant friction effaces the striae at short distances from the snout[53]; this was noted 100 m from the ice in Iceland and 300 m from the Lower Aar Glacier, and has been proved experimentally.[54] The facets too, though occasionally preserved, are generally lost by attrition.[55] The big angular blocks in the outwash, e.g. those of granite at Basel, of nummulitic limestone in the Mosbach Sands near Wiesbaden or of igneous and metamorphic rocks in Louisiana and in the Mississippi valley 200 miles (320 km) south of the nearest ice,[56] may have been obtained by floating ice or by fluvioglacial streams undermining faces of boulder-clay.[57] Similar large erratics were drifted to the neighbourhood of Vienna.[58]

The long, often narrow and shallow depressions, bounded by steep back slopes, which groove Pleistocene outwash,[59] have their counterparts in the waves which cross the Icelandic *Sandar*[60] parallel with each other and with the ice-edge. They represent the unfilled foss, noticed for example in front of the modern Alaskan glaciers,[61] between the low cones of two successive stages in the building up of the plain; they are associated with the "backset beds" that were made by later slumping or the upflow of water necessitated by deposition[62] (see p. 436). Thus the position of the ice-face may be indicated by an ice-contact slope with concave, steep and billowy surface and backset beds; by kettle-holes and kame-moraines and masses of till; and less clearly, by irregularly and poorly sorted material. The ice-contact slope has sometimes suffered postglacially or been steepened by subglacial streams when the ice withdrew.[63]

An oscillating ice-edge may make the surface undulate or sculpture grooves

or striae in it[64] (cf. p. 218). Certain dislocations and pressure disturbances[65] may, however, be made by river ice acting during the winter months when the outwash streams freeze over.

While the term "fluvioglacial" has been widely used, the term "glacifluvial" (*glacifluvium*) is probably to be preferred.[66] J. W. Gregory[67] introduced the term "glacieluvial" to connote the broad spreads of gravel laid down by irregular wash or sheets of water issuing from the ice-slope.

Valley trains. Chamberlin's "valley trains",[68] the "frontal deltas" of G. H. Stone[69] and "glacier-river formations" of Danish writers,[70] resemble other outwash plains in the great aggradation and coarseness of the material where their heads blend with moraines or outwash plains.[71] Unlike these, they are narrow and confined to valleys, filling these from side to side and to considerable depths—in the area of the Wisconsin glaciation to depths in places of as much as 90 m and in that of the Rhône Glacier to a depth of 120 m[72]—and spreading out as fans at the vomitory.[73] The decline down the valley which may be 20–50 ft (1 : 106–264) per mile and exceptionally as high as 375 ft (1 : 14) per mile is frequently not regular because they were deposited in sections corresponding with stages or halts in the retreat.

Active valley trains are highest along their middle line and slope laterally as well as down the valleys (though the lateral slope may later be reversed by compaction[74]). Aggradation, therefore, may form lakes or sluggish rivers in the lower courses of ice-free tributaries or in those which had less vigorous glacial drainage, or alternating lacustrine and fluvial phases of deposition. Such lakes had usually indefinite shore features since they were short lived and their level fluctuated considerably because high and low water in the main stream also had a big range. They were associated with valley trains[75] in the Alpine glaciation and about the North American ice-sheet: the tributaries of the Mississippi in south Illinois and of the lower Ohio were extensively ponded—silt, often beautifully laminated, covers the floors—and the gravels filling the Spokane Valley in Columbia formed, for example, Newman Lake and Liberty Lake.

Valley trains are found in almost all glaciated countries,[76] e.g. Scotland and the Caucasus, the broader valleys of north Germany (where they form the *Talsand* and some of the *Heidesand*), in north-west Russia and parts of the lateglacially superaquatic terrain of Fennoscandia, in eastern North America, e.g. along the Hudson, Delaware, Susquehanna and Ohio rivers, and south of the Great Lakes region, e.g. in the Wabash, Illinois, Wisconsin and Mississippi rivers, and along the Columbia River to the Pacific Ocean. They were often subsequently dissected as in the Lower Mississippi[77] which carried a great volume of melt-water from the ice-front; the overloaded debris from the ice, which had advanced into the basin and caused aggradation, was dissected by large streams discharging from glacier-lakes in which the sediment had settled. Dissection began immediately the ice withdrew and the zone of outwash deposition shifted and before the outlying gravel-buried masses of ice had melted.

Outwash, a series of cones. Outwash plains in countries of low relief may be one more or less continuous sheet, where the ice has advanced. If melting predominates low but distinct, flat semi-cones are built up,[78] concentric about the mouths of rivers issuing from subglacial tunnels and frequently in embayments in the ice-face. The apices are situated at the

breaches of the frontal moraines, as at the distal end of *Rinnenseen* (see p. 241), tunnel-valleys or föhrdes (see p. 353), the cones quickly broadening away from the ice and declining in height, at first somewhat rapidly, later more gently. The cones may lie between the lobes of the moraines or about their middle points. The first type (*Sanderbuchten*[79]; "interlobule fan"[80]), in which the emerging streams largely controlled the shape of the ice-front, occurred in Jutland[81]—it forms the *Hedesletter* west of the Main Stationary Line (*Hovedopholdslinie*)—in north Germany[82] (fig. 76) and in west Russia[83] where they have slopes of 3–5:1000 and the difference in height between

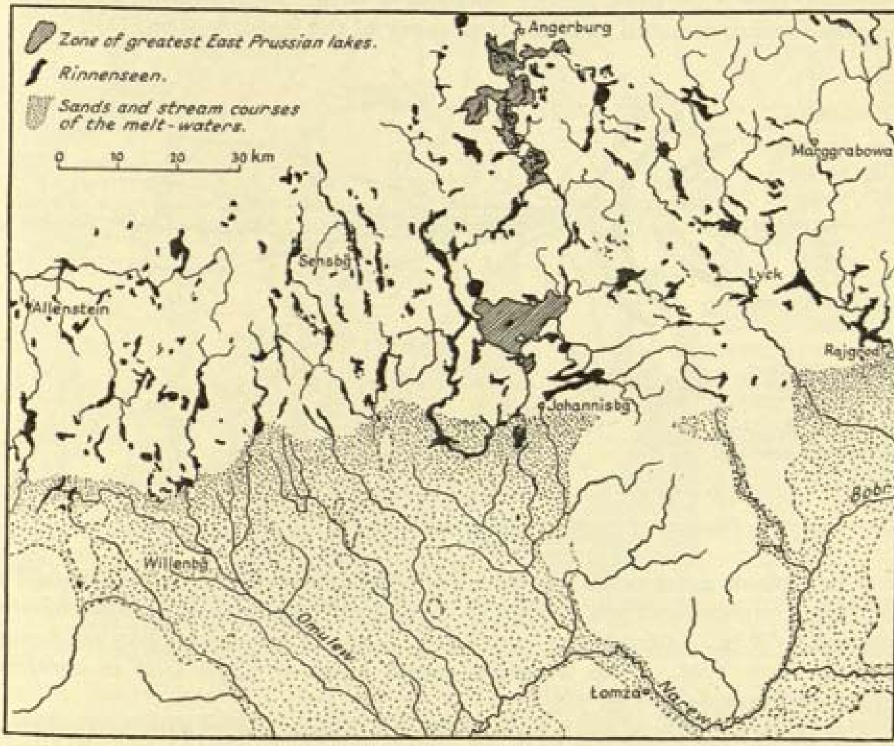

FIG. 76.—The East Prussian lakes and the sands of the Narew system. P. Woldstedt, *1820*, p. 117, fig. 11.

the floor of the lake and the highest part of the cone is 50 m, exceptionally 100 m. They sometimes display melt-water channels which carried off the overflow from lakes ponded during the recessional stages, as in the *Rinnenseen*.[84] The second type,[85] in which the ice apparently depended more upon the relief, was associated with piedmont glaciers north of the Alps. Grain size and current-bedding enable the drainage in both types to be reconstructed.[86]

Melt-waters may also cut down into the cones. Troll[87] has fully described their mechanism and the forms they assume on the Munich Plain. "Trumpet valleys" were excavated with new cones at the mouth (fig. 77) as soon as the retreat began, i.e. at the very end of the Glacial period and to a less extent in interglacial and postglacial times. Each recessional stage, due to the late-glacial dissection of the schotter and the formation of climatic meanders, had

its own trumpet valley and cone, the latter reaching farther and father outwards. Trumpet valleys occur in other valleys in central and western Europe, e.g. Alz, Inn, Lech, Moselle, Rhine and Ticino.

Pitted plains. Pitted plains (Swed. *kittelfält*) resemble other outwash plains in composition but are studded with kettle-holes, some of them so shallow that they hardly deserve the name. The pits have slopes of 5–20° and are circular, elongate or oval or quite irregular in shape. Usually they are unaccompanied by knobs though plains with few holes grade into others with countless hollows and finally into gravelly ridges and mounds of the kame-morainic type. Portions of the underlying drift, such as drumlins, osar, terminal moraines or till, are visible in many of the larger kettles.

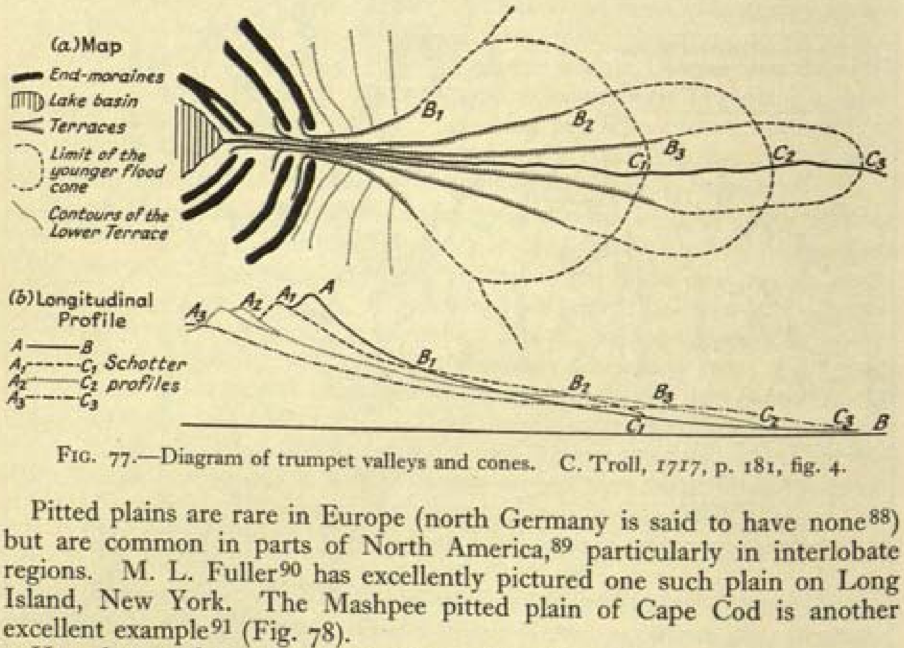

FIG. 77.—Diagram of trumpet valleys and cones. C. Troll, *1717*, p. 181, fig. 4.

Pitted plains are rare in Europe (north Germany is said to have none[88]) but are common in parts of North America,[89] particularly in interlobate regions. M. L. Fuller[90] has excellently pictured one such plain on Long Island, New York. The Mashpee pitted plain of Cape Cod is another excellent example[91] (Fig. 78).

How they are formed has been witnessed in Alaska[92] where the kettle-holes range in diameter from one or two metres to *c.* 75 m and number as many as 100 or more per square mile. The bigger ones have pools through which water flows continuously. Ice underneath keeps the waters cold and the slopes steep. It is responsible too for the constant slipping of the stratified gravels on the sides and the marginal cracking and faulting of the areas by slumping. Crumbling, infilling and meandering streams make the walls less steep: many kettle-holes are obliterated in this way. Extreme pitting implies the deposition of gravels over a nearly horizontal sheet of inert ice of irregular thickness.[93] These excessively pitted surfaces resemble kame-moraines but may be distinguished by their better assortment of material, their more even sky-line, and a tendency to lengthen along the valleys.[94]

Relation to moraines. Penck[95] and others assert that outwash plains are constantly associated with terminal moraines unless, as in narrow valleys, the glacial floods were big. O. Ampferer[96] contends that the two kinds of accumulation are inversely proportional and mutually exclusive. The Brandenburg and Frankfurt stages of north Germany (see p. 1164) are

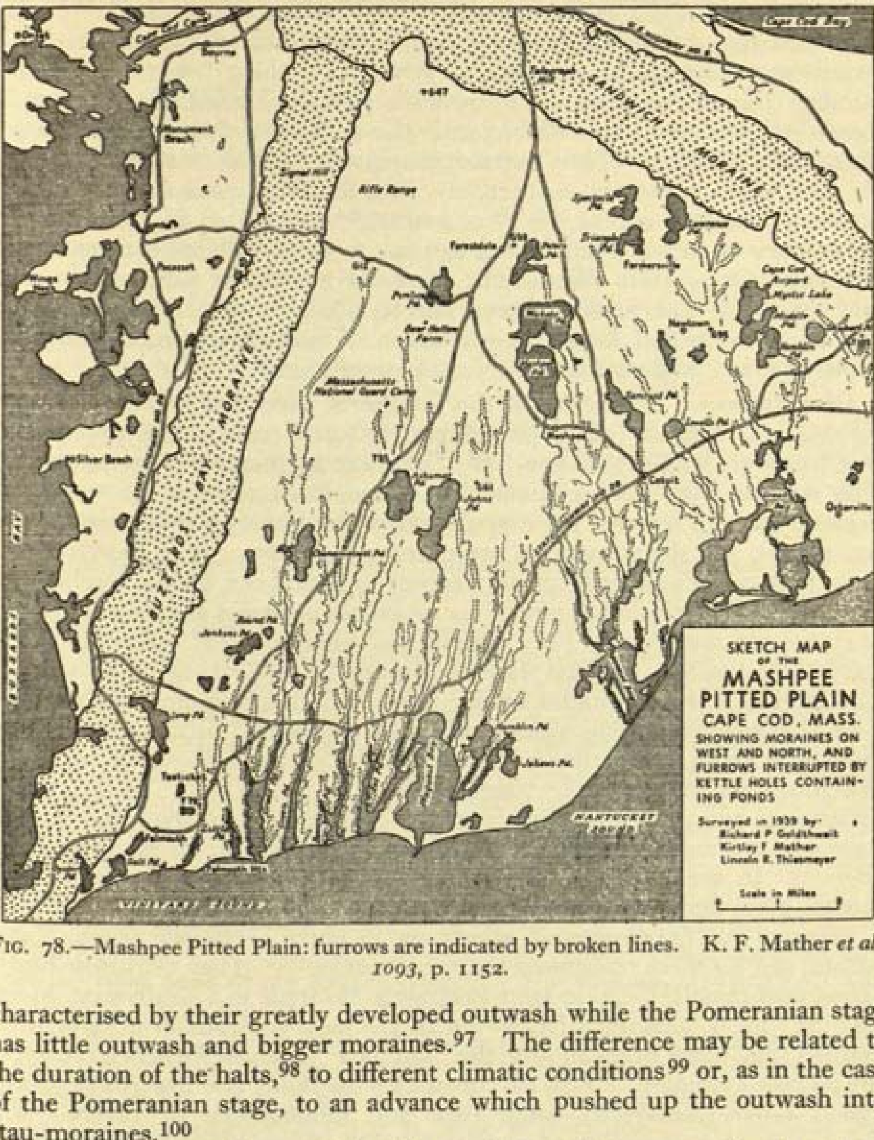

FIG. 78.—Mashpee Pitted Plain: furrows are indicated by broken lines. K. F. Mather *et al.*, *1093*, p. 1152.

characterised by their greatly developed outwash while the Pomeranian stage has little outwash and bigger moraines.[97] The difference may be related to the duration of the halts,[98] to different climatic conditions[99] or, as in the case of the Pomeranian stage, to an advance which pushed up the outwash into stau-moraines.[100]

The association of outwash and moraine, which occurs about modern glaciers,[101] is depicted in Penck's ideal glacier suite, the Glacial Series[102] or trinity (fig. 79; cf. also fig. 53, p. 270)—the other members are the *Zentralbecken* (see p. 269) and the drumlin zone (see p. 393)—and recognised in the

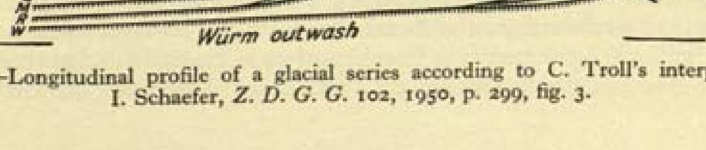

FIG. 79.—Longitudinal profile of a glacial series according to C. Troll's interpretation. I. Schaefer, *Z. D. G. G.* 102, 1950, p. 299, fig. 3.

Alpine glacialists' "cone of transition" or "region of passage".[103] The transition may be perfect, the outwash becoming steeper and coarser as it passes into the moraine, or there may be a low scarp between or a fosse parallel with the moraine where outwash partially buried the ice-front.[104] Sometimes detritus accumulated inside the moraine if this was massive and barred the drainage. Thus outwash may occur with or without moraines; the two types constitute respectively Chamberlin's overwash aprons and outwash plains or aprons.[105] Woodworth[106] recognised in New England three types of outwash: (*a*) frontal moraine terraces with ice-contact slopes; (*b*) esker-fans or small plains built up at the mouths of subglacial streams and connected with feeding eskers; and (*c*) wash-cones which slope sharply and have ice-contact faces. A succession of such fluvioglacial complexes showing imbricated structure may mark a recession.[107]

Distribution. All the ice of the ice-sheets, save that which was transformed into vapour, finally passed into water: "centuries of precipitated moisture were let loose at once".[108] This was augmented by summer rains. The streams were more copious than the modern ones in the same area because the cold reduced the evaporation, vegetation (which absorbs water) was absent, and the permafrost prevented percolation. It is, therefore, not surprising that Pleistocene streams appear to have been greatly swollen and that outwash sheets in their several forms should be one of the most important of glacial accumulations. For example, the "modified drift" extending westwards from the New England states to Minnesota, the Dakotas and Manitoba, was estimated at one-quarter or one-half of the total volume of the till.[109] The Mississippi outwash, which was fed by augmented precipitation and by melting from an 1800-mile (*c.* 3000 km) front of the Laurentide ice at the Wisconsin maximum, can be traced to the delta at the mouth of the river, a distance of 500 miles (800 km) from the confluence with the Ohio (a main outwash-carrying tributary) and 700 miles (*c.* 1120 km) from the nearest ice-edge.[110] In the Columbia River sediments of glacial origin reached its mouth, more than 300 miles (480 km) from the source of the melt-water.[111]

Some of the finest outwash sheets are in north Germany where Wahnschaffe and Keilhack recognised their true nature.[112] They constitute heaths (*Heidelandschaft*[113]), e.g. in west Schleswig-Holstein and south-west Mecklenburg; the Lüneburger Heide, one of the most unproductive parts of the country, and Mark Brandenburg, the "sandbox" (*Sandbüchse*) of Germany,[114] are notable examples. Outwash constitutes the sandy and gravelly platforms called *Geest* (*Sandgeest*), often bare and monotonous or covered with heather, bushes or woodland or with patches of bog (*Moorgeest*), which increase in area westwards and are often underlain by pans (*Ortstein*). The outwash consists of *Heidesand* or *Decksand* (Dut. *keisand*) and grades into bouldery deposits (*Geschiebesand*; *Geschiebedecksand*[115]) or into extremely fine clay devoid of sand and gravel (*Deckton*[116]; if banded, *Bänderton*). In the direction of the streams, it may be up to 65 km long, e.g. where the ground sloped from the Baltic Ridge to the Berlin *Urstromtal*.[117]

Other plains underlie the heaths of Jutland and Denmark[118] west of the Main Stationary Line, and fall at the rate of 1–2:1000, though their present freedom from trees is due to neolithic clearance of the open oak forest.[119] They extend westwards to the North Sea from the moraines which build the eastern half of the peninsula, as Forchhammer noticed,[120] and at their western margin give rise to the *vader* or *Vaderhavet* which are flooded at high

tide.[121] The heaths are now displaced by cultivated fields and plantations. The outwash plains occur too east of the Baltic[122] and in Russia,[123] e.g. down the Dnieper to the rapids at the mouth of Samara, and with fluvial sediments over much of Holland,[124] west of the Rhine (*Zanddiluvium*[125]; *Diluvium remanié*[126]). The relief and the widespread lateglacial submergence made them rare in Scandinavia, though they have been described, for example, from north Sweden (Frödin, 1954).

Outwash plains, up to hundreds of metres thick, are also found in the South American Cordillera[127] between 15° and 25° Lat.; and east of the South Alps of New Zealand, as in the Canterbury Plain.[128] This plain, 600 miles (*c.* 965 km) long and 30 miles (*c.* 48 km) broad, falls from 1500 ft (*c.* 450 m) to sea-level with a gradient of 1:117. Its vast coalescing outwash, 1800 ft (*c.* 550 m) thick, includes marine and lignite layers and fills in the shallow sea that once separated Banks Peninsula from the mainland.

In most regions, the outwash was probably built chiefly during the dissolution; for the lower temperatures, more rapid flow, and steeper fronts of the advancing stage provided a narrower zone of ablation and a correspondingly smaller discharge of melt-water.[129]

Alpine schotters. The Alpine schotters are among the most important of all outwash plains; with the loess they provide the basis of Pleistocene archaeology and chronology and the battle-ground of the controversy concerning the control of outwash plains. The subject of a considerable literature,[130] they are up to 70 m thick[131] and are widely distributed. As the *Terrassendiluvium* of the writers of the middle of the 19th century, they spread along the major valleys to their confluence in the Swiss Plain and down the Rhine to Basel[132] (the Rhine outwash is *c.* 75 km long), as well as over parts of Bavaria, as in the Münchener Ebene (Munich Plain)[133] which is 70 km long and 10–40 km broad between the Inn and Isar glaciers (fig. 80). They form big fans in Piedmont and Lombardy and over the site of the Pleistocene subsidence in the Po plain[134] where, as at Cremona, they are 237 m thick.[135]

The Rhine received melt-waters from all the Swiss glaciers, i.e. from the left flank of the Rhine Glacier, from the Linth, Reuss and Aare glaciers, and from the right flank of the Rhône Glacier. The various schotters joined at the confluence with the Aare and spread as a united outwash down the Rhine past Basel (fig. 81). The slope diminishes down the valleys and approaches more and more that of modern streams. It varies, however, within considerable limits[136]: the lowest is between 1·2 and 3:1000, the highest as much as 10–12:1000 and well exceeds that in north Europe (see above) because the valleys were narrow, the gradients high and the rivers big.[137]

Nature of control. Penck and Brückner[138] developed the view, which C. Martins[139] had expressed as early as 1841, that the schotters were built up fluvioglacially. Thus the schotters pass into moraines[140] though the dovetailing is seldom visible[141]; they enclose morainic material and are absent within the moraines,[142] though these are only to be expected in this position if the ice advanced over its own outwash[143]; their fauna, mainly larger mammalia (very rarely a microfauna), is arctic-alpine—the Deckenschotter has not yet yielded any vertebrate remains—and certain of their land shells are typical of the loess[144]; they are separated by loess which rests in hollows on their weathered surfaces[145] and by interglacial epochs of river erosion[146] during which they were either laid down higher up the valleys[147] or were not

deposited because the climate was too dry.[148] The Swiss rivers have cut down into the Lower Terrace to an average depth of 10–15 m.

This correlation has been fairly generally accepted[149] though glacialists differ about the exact chronology: the schotter formation is placed during the advance,[150] during the retreat,[151] or during the advance and retreat—two schotter terraces are so united into a single glaciation.[152]

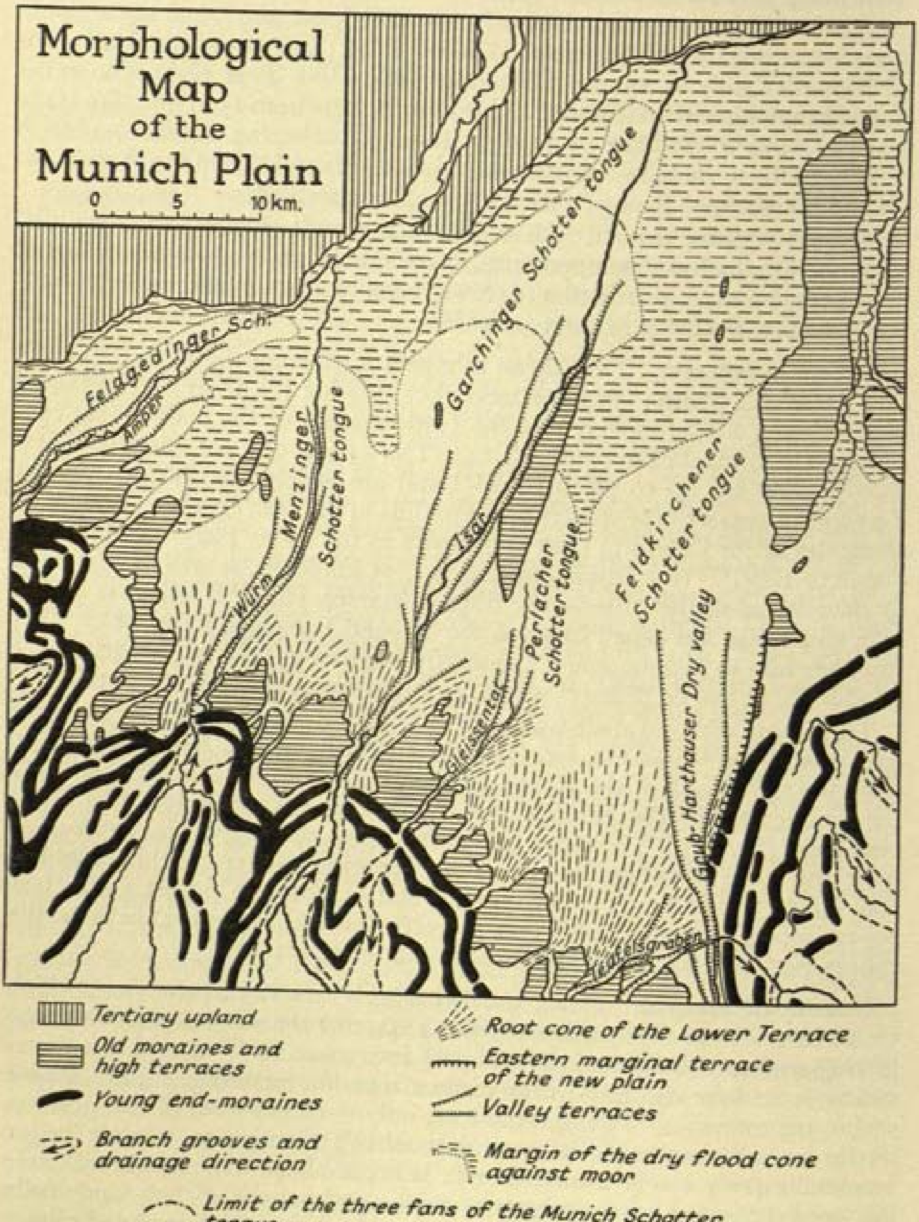

FIG. 80.—Morphological map of the Münchener Ebene (Munich Plain). C. Troll, *1717*, pl. I (opp. p. 172).

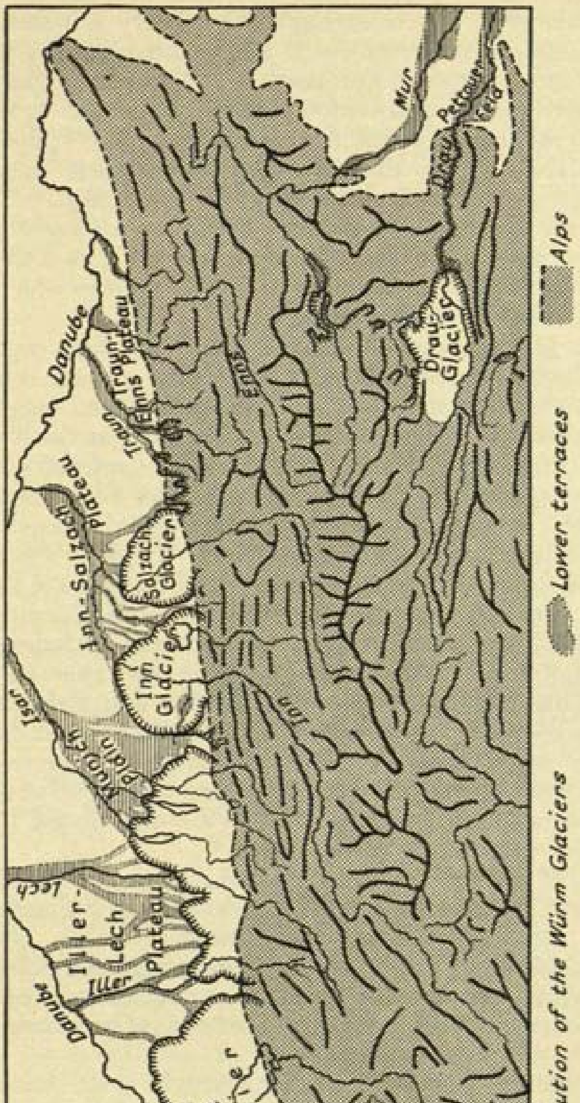

FIG. 81.—Relationship of the Lower Terrace to the schotterfields and glaciers of the Würm glaciation around the north side of the Alps. C. Troll, *1717*, p. 199, fig. 6.

Greater divergence is shown by those who think the schotter was induced by tectonic upheavals[153] which increased the delivery of material or, more generally, by tectonic subsidences which, during interglacial epochs, lowered the valleys and converted them into lakes.[154] With this origin, it is urged, are in better agreement the distribution of the precipitation[155]; the schotters' passage beneath moraines into the intermorainic region[156]; their furrowing by ice of the advance, e.g. in the Inn Valley[157]; and their independence of the end-moraine and the different ages of these deposits. The schotter formed during the advance has been termed the *Vorschotter*[158] or *Vorstoss-schotter*.[159] The position of the earlier moraines with its later schotter in Penck's classic diagram[160] is removed by J. Knauer[161] who suggested a readvance (W2) over W1 (cf. p. 1161).

This view, put forward originally by V. Hilber,[162] has been cogently argued by Ampferer[163] who showed to Penck's satisfaction that the schotter in the Inn valley was not laid down in a glacier-lake, as had been suggested,[164] but by a tectonic subsidence. The schotter extends as far as the Alpine margin and at Rum,[165] near Innsbruck, to 360 m below sea-level, giving a total thickness of over 450 m. The subsidence took place either continuously[166] or, more probably, with pauses denoted by the threefold alternation of coarse and fine sediments and the *Verlandungsserien*. Ampferer's reconstruction may be accepted for the Hochterrassenschotter[167] and for the Inn valley.[168] While Heim[169] attributes the drowning to a uniform subsidence of the Alps *en bloc* in relation to the foreland (see p. 280), others, including Penck and Ampferer, relate it to movements which varied in their amounts from valley to valley.[170] This bending, demonstrated for the Inn, Salzach, Enns and Drau,[171] for the Rhine,[172] for the valleys between the Inn and Linth,[173] and for the Isonzo,[174] made possible the great thicknesses of the interglacial lake-sediments (with their *Schieferkohle*) that have been discovered in these valleys, e.g. 200 m in the Lech, 400 m in the Ill and 300–700 m in the Inn.[175] The inner Alpine valleys have no such sediments.[176]

The existence of one widespread interglacial schotter, referred by Bayer[177] to the Riss-Würm but by others to the Mindel-Riss in Switzerland and the Riss-Würm in the eastern Alps, seems incontrovertible. Nevertheless, Penck,[178] while admitting the existence of an interglacial *Schotterunterbau* and of a fluvioglacial *Schotterunterbau*, remains with Heim[179] of the opinion that there are four glacial schotters as worked out in his classic publication (see p. 933).

The cause of the subsidence (see p. 280) is sought in ice-isostasy[180] or the compensatory waves and uplifts this produced outside,[181] or in a continuance of the Tertiary movements (*Grossbewegungen*) of like sign.[182] The latter is the more likely since sedimentation did not take place in certain big valleys,[183] e.g. upper Rhine, Tessin, Bergell and Zoisach, and the movements are difficult to reconcile with the prolonged lag in the uplift and with gravity compensation. Moreover, their local variations parallel those of Tertiary time, the Alps being still in a labile state as recent movements, for example in the Isonzo region,[184] testify.

The terraces in the Upper Rhine, Neckar and Main have also been equated with tectonic movements.[185]

A number of writers[186] invoke changes in the level of the ocean or of inland seas like the Sea of Aral[187] which shifted the basis of erosion. The terraces were controlled by synchronous oscillations, chiefly negative movements

interrupted by positive ones of less amplitude (see p. 1268). This view fails, however, to take into account the schotters' dovetailing with the moraines and their downstream convergence.[188] Hence, a single valley had two controls, a glacial and fluvioglacial one above, and an oscillating sea-level below,[189] as the grading of the outwash with fluviatile terraces suggests.[190] The same glaciation produced fluvioglacial terraces in the high valley and rejuvenation in the lower valley following the eustatic lowering of the sea (see p. 1355); interglacial times witnessed erosion above and aggradation in the lower parts[191] (see p. 1026). Terrestrial movements in the middle stretches, as in Rhine and Danube (see p. 1265), have complicated matters.

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Chapter XXIII

GLACIO-LACUSTRINE AND GLACIO-MARINE DEPOSITS

1. Glacio-lacustrine Features

(a) *Characteristics*

Lakes on modern glaciers. Under continued thaw, glacier-surfaces which have a gentle, even fall and are comparatively free from strain, become seamed with stream courses, sometimes unfordable, and dotted with pools and lakes which may be formed by the collapse of englacial or subglacial passages, by enlargement of parts of crevasses by melting, or by the more rapid melting of clean ice amidst debris-protected ice. The lakes travel with the ice and drain through fissures or over low points on their rims. Superglacial lakes of this kind have been observed in the Alps[1] on the Mer de Glace, Géant, Gorner and Aar glaciers. They occur too in the Karakoram Mountains and Himalayas,[2] in Alaska,[3] on Vatnajökull,[4] and in North-East Land during summer[5]—in striking contrast with the spring as A. E. Nordenskiöld's traverse showed.[6] Streams, with lakes up to 1 km long and 5 m deep, are also to be found on the Greenland ice-sheet,[7] particularly in the west where they are met with up to 1530 m above sea-level and 107 km from the ice-edge.

In the Antarctic, the warm arctic summer has no parallel. Because of the extreme cold (see p. 167), dust wells are few,[8] crevasses are deeper and remain wide and gaping,[9] and the sea-ice keeps its normal salinity up to the ice-edge.[10] Melting is local and infrequent and flowing water is almost completely absent.[11] Yet streams and lakes are not unknown,[12] especially on northern faces, or where strong foehn melts the snows along the inner margin of Ross Barrier or rock-walled valleys encase the glaciers, e.g. the Ferrar Glacier. Superglacial lakes repose towards the periphery on piedmont glaciers of gentle slope, notably if the ice is nearly inert and therefore uncrevassed, or is so cold that streams are compelled to remain superglacial.[13] In the interior[14] (as well as in Greenland[15] and on Mount Everest above 7600 m[16]), there is no melting, no surface stream, no hard snow crust, and no macroscopic stratification of the firn.

Nunatak lakes. Lakes are ensconced in the lee of nunataks[17] ("lee-lakes"[18]) or of projections into a glacier's side, and in the "moats" of Hobbs[19] which are sometimes bridged with snowdrifts or converted into the courses of glacier-streams.[20] Usually deep and narrow and hemmed in by vertical faces of ice, they owe their origin in a small measure to the wind, mainly to absorption by the rocks of the sun's heat in summer and its radiation which melts the ice back.[21] Waters streaming in rivulets from the ice either flow over it or disappear through ice-grottos.

Such lakes are not infrequent in Greenland,[22] in both east and west, as around the Jensen's Nunataks.[23] They are rare where the ice-surface in east Greenland mounts towards the nunataks[24] because here the flow is more active or greater nourishment replaces loss by ablation. They are extremely few on

the Antarctic continent but have been seen in South Victoria Land[25] and round Gaussberg whose deficient precipitation fails to compensate for the local ablation.[26]

Pleistocene nunataks below the snowline were also associated with lakes. Their traces are usually indistinct because they were ephemeral, the subglacial drainage varied their water-levels, and their sands and gravels escaped into subglacial or lateral streams or were afterwards modified by solifluxion and ice-action. Nevertheless, lines of pebbles and strandlines margin the nunataks[27] in south Finland and Greenland though it is sometimes difficult to distinguish them from horizontal moraines or fluvioglacial accumulations.[28]

Modern marginal lakes. Lakes are hemmed in along the margins of glaciers under special conditions of restraint. Of the five recognizable varieties the most important is found where a main glacier, advancing down its valley, closes its ice-free tributary. The closure is only effective if the trunk valley is not much or at all overdeepened, or if both trunk and tributary have been similarly lowered. This is the commonest type in the Alps[29]: the Oetztal, Rutor and Gurgler lakes are or have been instances. The classic example is the Merjelen See[30] which is bounded by a tongue protruding eastwards from the Aletsch Glacier. This triangular lake, which is at 2435 m and is 78·5 m deep, attained its greatest length of 1600 m in 1878. Graenalón,[31] the largest glacier-lake in Iceland and one of the longest in the world, has an area of 188 sq. km and a depth of 200 m. Marginal lakes are also ponded in Norway[32] (Demmevatn is the best known), Iceland,[33] the Pamirs,[34] Himalayas and Karakoram Mountains,[35] Tienshan,[36] west and south-west Greenland,[37] including those in side-fjords that rise and fall with the tides.[38] They are less common in north and east Greenland,[39] in Baffin Land,[40] in Spitsbergen,[41] in the Caucasus,[42] in British Columbia,[43] in South America[44]—Lago Argentino has a surface of 1560 sq. km—and very rare in the Antarctic.[45] Alaska[46] has the most abundant lakes of this class.

The second type, which arises when a lateral glacier obstructs the main drainage, is seldom observed, partly because, as is seen in modern glacier-regions (see p. 63) and has been inferred for the Pleistocene Thames,[47] the stream bars further progress by melting the ice and tearing off floes. Instances include the Mattmarksee held up in the Saas valley by the Allalin Glacier[48] (1817); the lake in the Rofental ponded by the Vernagtferner at times of great advance[49] (1848); the Lac du Combal at the southern base of Mont Blanc; and the lake in the Shyok valley of north-west India[50] (see below). In the Karakoram Mountains,[51] where the type is not uncommon, one such impounding glacier carries the illuminating native name of Chhatiboi[52] ("there will be a lake"). This relationship was very common in the North American Cordillera in Wisconsin time: an example is the Spokane River Valley and its continuation, the Columbia River valley as far downstream as the Coulée Dam.[53]

A third and rare type[54] is the triangular, usually shallow and diminutive lake which nestles at the junction of two glaciers because of their convexity. Alpine illustrations[55] are the Lac du Tacul at the end of the spur of Mont Tacul, the lakes between the two branches of the Gorner Glacier below Monte Rosa, the lake between the Hintereisferner and Kesselwand, and that on the Glacier du Talèfre below the Jardin. Other examples have been recorded from the Himalayas[56] and the end of Dalager's Nunatak, west Greenland.[57]

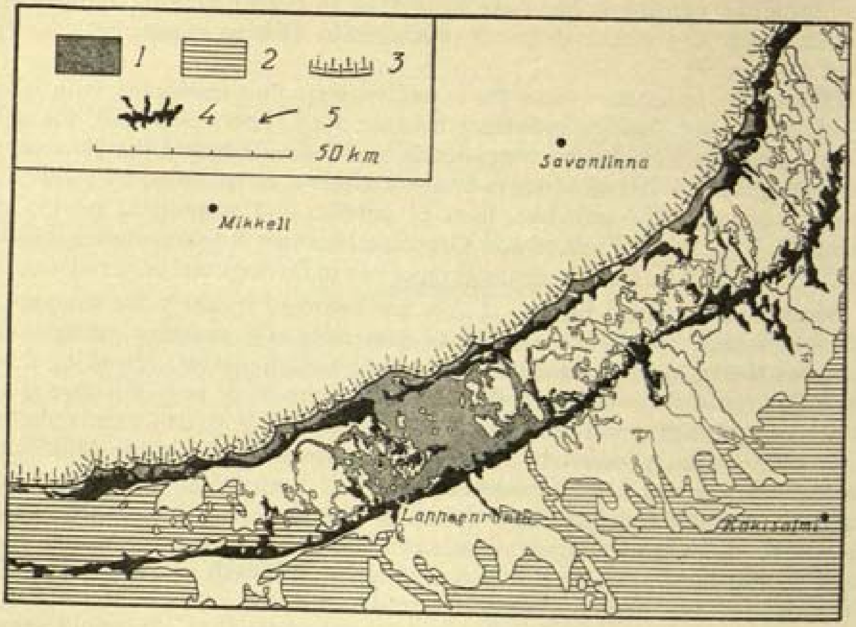


Fig. 82.—The early stage of the glacier-lakes of Saimaa and Sääminki in the region of the Salpausselkäs, Finland. 1, glacier-lakes; in the middle Saimaa, in the north-east Sääminki; 2, Yoldia Sea; 3, ice-edge; 4, the Salpausselkäs and a few osar; 5, overflow channel of the glacier-lakes. A. Hellaakoski, *F.* 59 (4), 1934, p. 82, fig. 19.

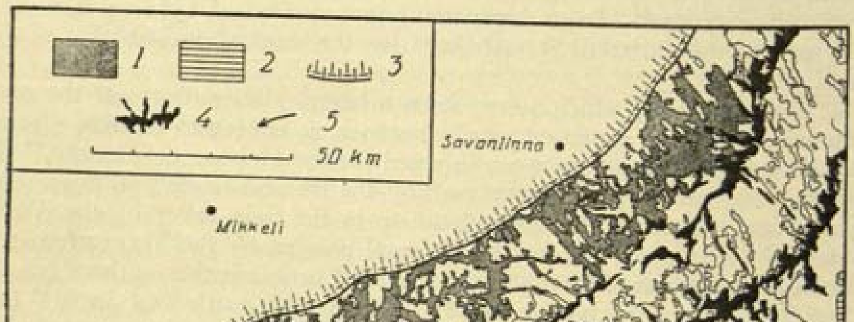


Fig. 83.—Second stage of the glacier-lake of Saimaa, Finland. A. Hellaakoski, *F.* 59 (4), 1934, p. 84, fig. 20. [Key as in Fig. 82.]

The remaining types embrace lakes which exist by double instead of simple closure—Lake Tasersuak[58] ("big lake") in west Greenland is the best known—and those which occasionally arise in a main valley by a glacier's retreat from its head owing either to ice-starvation, as in Alaska,[59] or to ponding by a remanié glacier, as in the Val de Bagnes below the Giétroz Glacier.[60]

Lakes ("melt-water lakes"[61]) are sometimes imprisoned between an ice-front and an earlier moraine spanning a valley through which glacier streams have been unable to keep a clear passage.[62] They are generally short-lived but were probably of some prevalence during the Glacial period[63]: the glacier-lakes of Saimaa and Sääminki which were held up by the Salpausselkäs (see p. 1172) are excellent examples (figs. 82, 83). Their waters either escaped through the sands and gravels, leaving their muds behind as banded clays,[64] or discharged over their moraines in hollows or "creases",[65] building up outside high, fan-shaped deltas, fans or valley trains.

Outbursts. Where the summer is warm enough to induce much melting, as in the Alps, Himalayas, Iceland, Spitsbergen, Alaska and South America, glacier-lakes are liable to drain completely or partly through or under the ice and flood the valleys below. They give rise to submarine springs in fjords, or in lower glacier-lakes to sublacustrine springs which sometimes spurt up with great volume and rapidly empty higher lakes.[66] This sudden evacuation of lateral, possibly also of englacial or subglacial lakes (see below), takes place by opening new or widening old crevasses by the melting and undermining action of glacial waters, by escape between a glacier and its bed, or by spouting out as fountains on its surface or through subglacial sands and gravels.[67] The drainage starts as a mere trickle along the crevasses but rapidly widens these into tunnels which enable the flood to break through.[68] The volume of such waters may be enormous. It is estimated,[69] for example, that the burst of the Giétroz Glacier in 1818 discharged 1110 million gallons, the Merjelen See in 1878 1709 million gallons in 9 hours, the Shyok Glacier (see below) in 1929, 1475 million cu. m and the Grimsvötn eruption in 1938 50,000 cu. m/sec or a total of 7·5 cu. km. The most violent floods in Norway occur when the glacier begins to float and the waters form a wide expanse beneath it[70] (fig. 84).

Statistics and historical records from the modern Alps[71] show that catastrophic outbursts of this kind are a normal feature of ice-activity though the present recession (see p. 146) makes them less frequent. Thus the Mattmarksee has burst out 26 times since 1859[72] and the Merjelen See[73] discharges at irregular intervals, associated with ice-recessions—it was drained in 1939 and again in 1947. The bulk or whole of the water passes through fissures in the Aletsch Glacier into the Rhône valley, lowering the lake-level at first slowly, then more rapidly, and exposing a vertical face 500 m long. The burst sweeps as a wave, which on the 18th and 19th of July, 1878, carried 11 million cu. m of water in 30½ hours. To reduce the flooding, a tunnel 548 m long was constructed in 1889 from the lake into the Seebach; the retiral of the Aletsch Glacier has up to the present made it unnecessary.

Similar outbursts have been noted in the Himalayas[74]—the Shyok Glacier[75] on the Indus produced the most important, its impounded lake achieving a surface area of 65 sq. km and a depth of 122 m (fig. 85). On bursting it generated a wave which rose to 26 m, attained a speed of 13·7 miles (22 km) per hour, and transported blocks of ice up to 15 m cube. Others have been witnessed in western North America,[76] in Scandinavia[77] and in the Mendoza

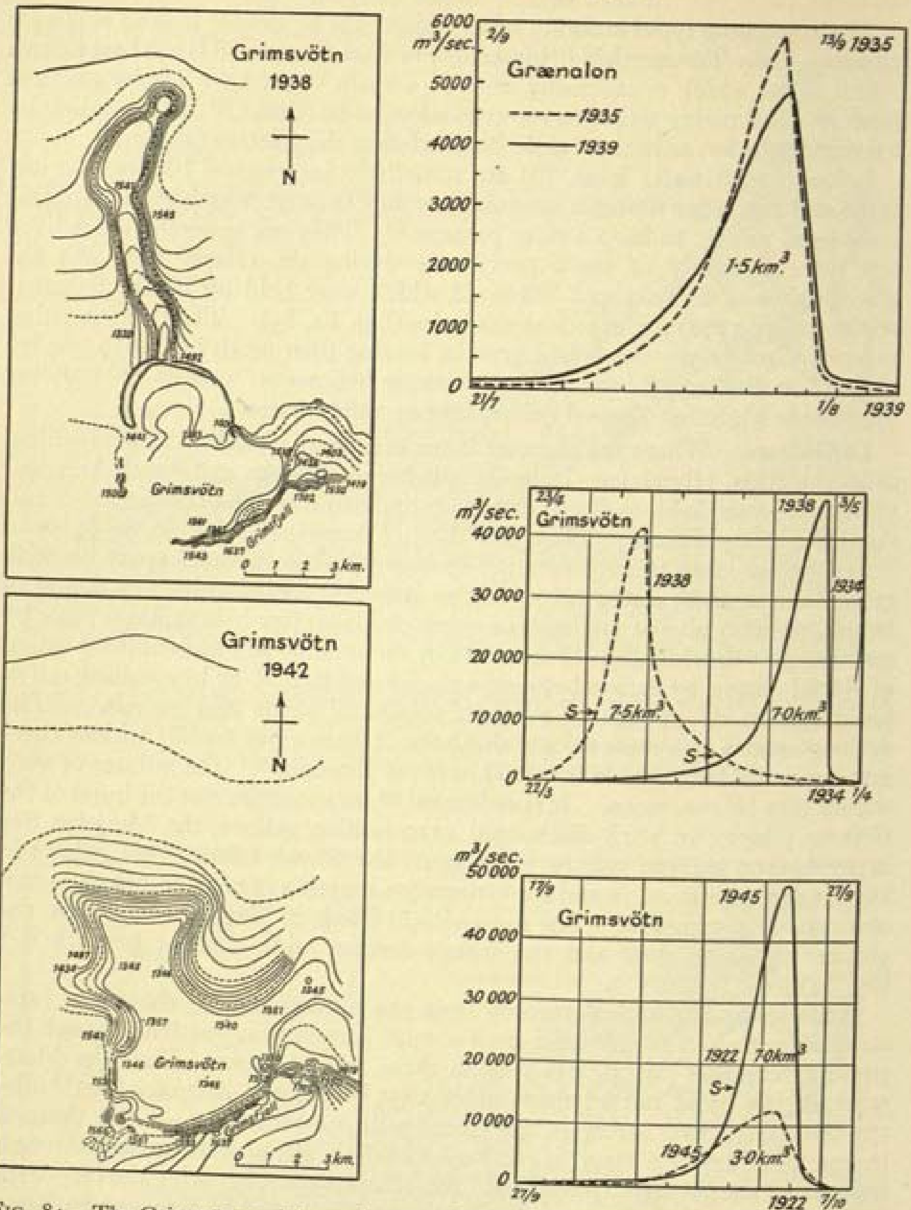


FIG. 84.—The Grimsvötn and the outbursts. *Top left*, Grimsvötn depression and the area north of it after the 1938 outburst (heights in metres, contours at 20-m intervals); bottom left, Grimsvötn in 1942; *top right*, approximate discharge graphs of the outbursts from Graenalón in 1935 and 1939; *centre right*, approximate discharge graphs from Grimsvötn in 1934 and 1938; *bottom right*, approximate discharge graphs from Grimsvötn in 1922 and 1945. S. Thorarinsson, *J. Gl.* 2, 1953, p. 271, figs. 4–10.

valley (Plomo Glacier) and other valleys of South America.[78] In Iceland,[79] where Grimsvötn bursts occur about every 10 years (e.g. 1861, 1873, 1883, 1892, 1903, 1913, 1922, 1934, 1945), the glacier outbreak or *Jökullhlaup* ("glacier-run") arises not only from the bursting of glacier-lakes but from melting by volcanic activity (see p. 65) and is heralded by an unusually high

silt content in the glacier-river. The flood may be so powerful that it floats the whole edge of the vast ice-cap the moment before the catastrophic break takes place. It washes away or levels long lengths of moraines and sweeps over the *Sandar*, carrying with it large amounts of mud and a mixture of dead ice, erratics and volcanic ashes.[80]

Subglacial and englacial lakes. Subglacial or englacial bodies of water, wholly ice-bound or partially confined in hollows in the ground, are associated with some glaciers. They are demonstrated by catastrophic emptyings (most likely from englacial sources[81]), as in connexion with the Ferpècle Glacier[82] or the Malaspina Glacier,[83] or in the catastrophe of St. Gervais-les-Bains[84]

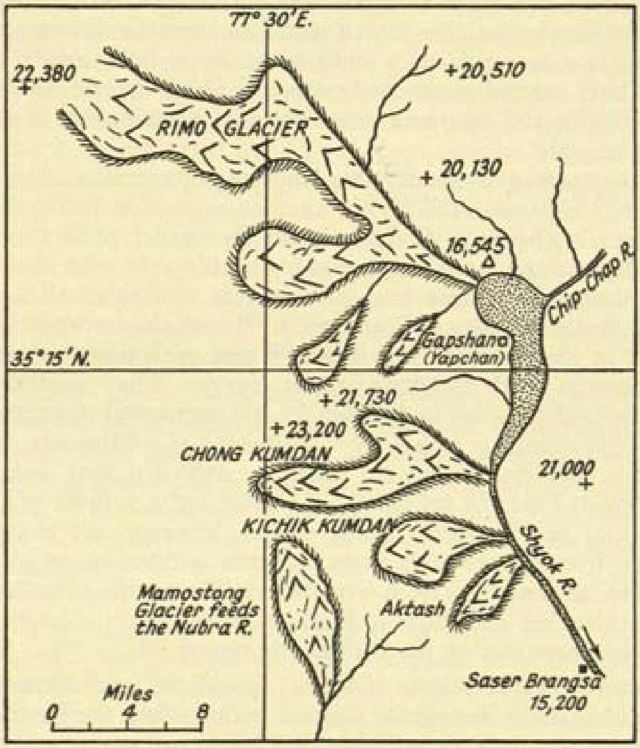


FIG. 85.—Sketch-map of the Shyok Glacier and the Shyok River. J. M. Lacey, *Engineer* 154, 1932, p. 372, fig. 2.

where waters from the Tête Rousse Glacier escaped from a gallery of a glacier-stream or from a lake completely enclosed. They explain the crater-like depressions on the ice after the escape of the flood,[85] e.g. on the Übelthalferner, Tête Rousse Glacier and Tasman Glacier, New Zealand, and account too for typical fault-lines, as on the Jostedalsbrae,[86] and for the clear glacier-streams, with regular régime, as in the case of the Puntaiglas Glacier.[87] Similar subglacial lakes have been suggested for modern Iceland[88] and for the Pleistocene ice-sheets (see p. 111), e.g. iceward of transverse ridges.[89]

Protection against outbursts is sought in the Alps by tunnelling under the glacier, e.g. the Allalin, or by constructing dams to check the waters, e.g. in the Martell valley of south Tyrol.

Geological action of outbursts. The destructive action of glacier-ice is considerable[90]: it includes the overriding of cultural features, etc., during glacier-advances (see p. 141); the action of avalanches (see p. 579); the bursting of subglacial or englacial lakes, of lakes between glaciers and earlier frontal moraines,[91] such as the outbursts in the Peruvian Andes where the level of water in narrow gorges rose 140 ft (*c.* 43 m)—about 6000 lives were lost in one such catastrophe—and of lakes held up by ice-avalanches[92]; and the sudden emptying of marginal lakes.

The geological importance of outbursts was early recognized.[93] Besides breaking off masses from the end of a glacier,[94] they temporarily augment the volume and therefore the erosion of subglacial streams, sweep away frontal moraines, screes and other materials from a valley's sides,[95] and build up torrential deposits by their "waves of sand", as from the discharging Merjelen See,[96] unloading as much as 1 million cu. m at one time[97] (see above). They may have carried some of the more outlying erratic blocks, as in the Punjab[98]; blocks 200 cu. m in size were conveyed 12 km during the St. Gervais catastrophe.

Periodic or occasional catastrophic emptyings were associated with Pleistocene lakes,[99] in Europe and North America, especially during the initial and closing stages. The geological results of the earlier phases are no longer decipherable, and the later ones are recognizable only with doubt and difficulty. Outbursts have been credited with the erosion of all our valleys,[100] with certain erosional signs in Scandinavia,[101] with the transport of numerous big erratics in the Durance and Etsch[102] and with unsorted and tumbled accumulations in north Sweden (Frödin, 1954). They washed the valley-walls and ravined hillsides in Sweden[103] and registered the extremely thick varves or "Jökullhlaup layers".[104] Glacial Lake Missoula in Montana, which had a capacity of 500 cu. miles (*c.* 2080 cu. km), escaped with a velocity estimated at *c.* 65 ft (20 m) per second and a volume of 9·4 cu. miles (*c.* 40 cu. km) per hour.[105] These effects, however, are obviously indistinguishable from those which arise from the sudden or rapid bursting of detrital dams, as described from Scandinavia[106] and the Himalayas,[107] or of ice-jams, avalanches and landslides which occurred on a large scale in the Alps and elsewhere during the Pleistocene retreat.[108]

Pleistocene glacier-lakes. Glacier- ("proglacial" or "hyper-glacial"[109]) lakes were ubiquitous during the Glacial period when the ice flowed against the drainage, e.g. south of the Baltic depression in Europe and of the Great Lakes in North America. They were only rare if, as within the bounds of the Alpine glaciation, ice and water moved in the same direction and down the natural slope. They were prone to lie on southern sides of valleys in the northern hemisphere and to be absent from northern ones, e.g. in the Aire valley of Yorkshire.[110] As the ice-margin receded and new outlets were opened and deepened, the lakes continually changed their outline, area and altitude. Immensely bigger than those of to-day, especially in regions of small relief, they enlarged and shrank, rose and fell, combined and parted in a highly kaleidoscopic manner in obedience to the topography. They were maintained by direct precipitation, by drainage from the surrounding slopes, by melt-water from the ice-edge and by outflow from higher lakes.

The ice-face, probably rendered steep by marginal melting, as in the present Merjelen See, was usually bowed outwards by the flow towards the lake,[111] though hydrostatic pressure prevented the ice from advancing too far.[112]

Melting and waves, combined with a calving of small bergs, sometimes associated with seismic action, forced the face in places to fall back more quickly than elsewhere and become concave, so forming "calving bays" as in Scandinavia[113] and North America.[114]

Exception has at times been taken to the theory of Pleistocene lakes of the dimensions generally postulated. For example, the depth assumed, such as the *c.* 460 m at the ice-face in Montana,[115] would lift the barrier and allow a subglacial escape[116]; the sites were occupied by stagnant ice, the supposed lacustrine clays being "pressed clays" or the banded dirt of englacial detritus, released by rising bottom melt[117]; and normal streams lowering their beds left the alleged lake-terraces.[118] The spillways originated before the Ice Age by river capture or by erosion along joint systems[119] (see below) and could not have been cut by the clear streams that issue from glacier-lakes.[120] But the quantities of debris in glacier-lakes and the thin winter-varves (see p. 1152) make it likely that much fine rock-flour went through the outlets, as it does to-day,[121] some of the streams in modern Alaska and Greenland being dirty brown or milky with sediment; the bare rock, exposed to frost, as in the high Andes, facilitated effective erosion; material lay ready to hand if the stream flowed over drift[122]; and ice-jams at the exit, with the irresistible force of the momentum of a flood, caused a rush of water when they burst at the spring thaw, the waters excavating rapidly and using the detritus which the winter frosts had accumulated.[123] Deposits beyond the spillways include rock-fragments of known derivation which could have reached their present positions only by coming through the spillways.[124]

Azoic. Modern glacier-lakes, as in Switzerland,[125] are generally azoic; they contain neither water plants nor animals, though the waters, if clear, may have a moderately rich biota.[126] A few species of copepods have been discovered in those of north-east Greenland,[127] some insect larvae in those of west Greenland,[128] and some life in those of the Yukon.[129] Glacier-lakes in Alaska contain ice-worms[130] and seals persist for many years in the glacier-lake in the tributary of the Kangerdlukasik ("peculiar fjord"), west Greenland.[131]

This sterility springs from several causes: the lakes are temporary; they lack organic matter which might serve as food; and their abundant detritus prevents a flora developing, as in the muddy stretches of the Greenland fjords, though Limnaeas are plentiful in Alaskan kettle-hole lakes which glacier-streams charge with fine mud.[132] The main cause may be the extreme coldness of the water, due to floating ice, to contact with the glacier and to the cold ice-winds, though cold glacier-streams are not without life.[133] The lakes belong to the polar type[134]; they have a permanent thermocline which limits the circulation to the epilimnion whose temperature does not exceed 3°C.[135] (see p. 1153). Pleistocene lakes probably froze over during winter, as they do in present-day Antarctica[136] and the Alps—the Merjelen See is frozen over 270–300 days in the year: this was surmised for Lake Agassiz,[137] though the absence of boulder-ridges from many stretches of its beaches may mean that it was not wholly ice-covered.[138] Freezing took place near the ice-edge where the stratification was inverse and the cold water on top.

Traces of life, apart from the occasional plants which have been swept into the lakes,[139] are not quite unknown from Pleistocene glacier-lakes : the plants from glacier-lake deposits of Jerseyan (?) age in Pennsylvania, which include *Betula nigra*, *Quercus predigitata*, *Castanea pumila* and *Ulmus racemosa*, might

be found in the same general area to-day. Tracks of *Chironomus* larvae have been yielded by Swedish lake-clays[140] and intermorainic clays (they may be interglacial[141] or interoscillational[142]), and indications of annelids, crustacea and fish by other lake-clays.[143] Additional instances of life have been recorded from Germany[144] (freshwater shells in the East Prussian drift were thought to be relict from such lakes[145]) where they are confined to the summer varves. Arctic plants have been found in similar clays either alone, as in Sweden,[146] or with freshwater molluscs, e.g. several *Planorbis* species in Fyn,[147] or with gastropods and ostracods in Poland.[148] *Silurus glanis* occurred in the clays of the Mitau ice-lake[149] and *Coregonus* in the Fish Lake of Leningrad (see p. 1292). The present uniformity of the fish fauna of the rivers of Europe between the Rhine and the Neva has been attributed to the interconnexions between the rivers as the ice withdrew.[150] Freshwater diatoms have been found in the Glen Roy terraces,[151] though they may presumably be modern.

Quite a number of specimens have been gathered from lacustrine sediments in the Pleistocene lakes of North America[152] which, to judge from fossil evidence and from modern distributive patterns of aquatic forms,[153] were only semi-glacial during their later history. They include shells from Lake Agassiz[154] (this shallow lake, its shores far from the ice-face, was readily accessible to species ascending from the Mississippi), freshwater shells from the Iroquois beach near Toronto,[155] with remains of mammals,[156] e.g. mammoth, elk, beaver and musk-ox, and plants,[157] and horns of elk and deer from the Leipzig beach.[158] Freshwater shells have also been extracted from the beaches of Lake Chicago[159] (with mammoth teeth), from the Algonquin beach[160] near Port Huron and elsewhere, from the Nipissing beaches[161] (with plants) at various places, and from Lake Ojibway.[162]

Drainage channels. Evidence of vanished glacier-lakes is provided by "fossil" drainage channels trenching main watersheds or projecting spurs and scoring hillsides or escarpments, as well as by deserted cliffs, beaches, deltas, floor deposits and marginal moraines. The water plane was controlled by the height of the bounding watershed and of the ice-barrier. It may be restored from the altitude of kame-terraces, beaches and deltas (these, after allowing for later settling,[163] fix the altitude of lakes that were too ephemeral to fashion other features such as cliffs, bars or terraces), from the intake of the channel carrying off the surplus water (when the small depth of water in the outlet is allowed for), and from the outfall of any feeder of the lake.

Just as lakes to-day may, as on watersheds, have two or even as many as five outlets working simultaneously[164]—they were probably common post-glacially in Scandinavia (see p. 1317)—so two or more outlets sometimes operated at the same time in connexion with Pleistocene glacier-lakes,[165] notably where hard rocks hindered downcutting.

Glacier-lakes may drain over the ice, through englacial fissures, or under the sole[166] (the usual mode in small glaciers as in the Alps): the subglacial chutes are sometimes still visible and cut across the system of marginal channels.[167] Waters which flow over glacier dams have a small hydrostatic pressure and erode mechanically and by melting. Basal drainage may be by basal crevasses, by permeable subglacial sediments or occasionally by the lifting of the glacier by the buoyant action of the water. The lakes may also discharge along the side, as is customary where the waters, impounded by big glaciers, have not ready access to subglacial channels.[168] Such marginal

drainage, which is usually short and of small volume, is exhibited about modern glaciers, as in Alaska.[169] The waters are either clear or greenish-yellow and turbid with glacier meal.

The overflow channels, the spillways of North America (Scandinavia[170]: *israndrännor*, *strömrännor*, *skvalrännor*, *daljor* or *torrdalar*), are unmistakable: they are indeed among the most impressive memorials of the Ice Age. Although possessing all the features which distinguish the work of streams, they are usually streamless (hence often termed dry valleys) but sometimes have disproportionately small streams or rivulets—the River Spree is an example of an underfit stream. They are excavated in rock or drift. Those in drift may present boulder-strewn surfaces or pavements,[171] not very unlike De Geer's boulder-channels[172] where melting rills crossing moraines have removed the finer material. Such ravines are frequently very deep. Thus the Grand Coulée[173] of Columbia, U.S.A., a canyon 50 miles (80 km) long, 1–5 miles (*c.* 1·5–8 km) wide, and 800 ft (*c.* 240 m) deep, has involved the excavation, during Illinoian time, of 40 cu. miles (*c.* 166 cu. km) of rock—this was formed by the headward retreat of a gigantic cataract, 800–900 ft (244–274 m) high and up to 5 miles (8 km) wide, formed by the overflow from a large lake ponded in the Columbia valley by an ice-lobe moving from Canada (cf. p. 240).

The channels have sharp edges at top and bottom. This lack of bevelling of the rims differentiates them from preglacial rejuvenation gorges, though glacial drainage may have utilised and modified the latter. The valleys are also typically flat bottomed and in their uniform cross-sections resemble railway cuttings (pl. XVIII A & B, p. 465). Those which fall steeply are often straight and immature and show abandoned waterfalls, cataracts, plunge basins and pot-holes and in limestones even caves.[174] When the gradient is gentle, they may have small tarns or marshes. They exhibit wide swinging curves which conform with the actual banks of their accurately proportioned streams. Terraces testify to the cutting away of successive obstacles.[175]

The intake, which has a very low fall and is often marshy and peaty ("swamp col"[176]), has frequently been scooped out by the swirl of convergent waters,[177] so that the channel falls outwards at its mouth. This sometimes makes it difficult in short valleys to ascertain the direction of flow. The outward fall at the head has also been attributed to postglacial stream action or to reversal of river-flow.[178]

All gradations are found down to small and indefinite scourways of short-lived streams from transient lakes. Yet the channels are frequently capacious canyons, especially if the rocks were soft, the withdrawal was brisk and the distance to an alternative escape was considerable. Their size was in keeping with the lakes they drained. In conjunction with their striking situation on main watersheds, their sudden termination on valley sides, and their narrow mouths, it made interpretation difficult: they have been regarded as local fractures or joint structures,[179] as relics of ancient water courses connected with river capture[180] or superimposed drainage,[181] as glacial grooves[182] or as valleys notched by a glacial sea into a rising land.[183]

The precise course the overflow waters took depended upon the slope of the ice-free land and upon the orientation of available routes, e.g. pre-existing valleys and cols. Four types are recognisable[184] (fig. 86): (*a*) direct channels across cols in a main divide such as the "col gullies" (Swed. *sadelskäror*) in the Scandinavia divide—they went out of action when the ice withdrawing from

a watershed opened up lateral communications; (*b*) submarginal channels across spurs of tributary watersheds or loops in deep, winding valleys[185]; (*c*) marginal channels following closely the edge of the ice and slanting inconsequently across the strongest line of fall; and (*d*) lateral gutters leading from the ice astride a water parting and sometimes spreading "morainic fans" down the hillsides[186]—the gutters developed into direct overflows where their streams on withdrawal of the ice continued to cut back steadily through the crest. The absence or unexpected smallness of the channels in an otherwise complete sequence of overflows is probably due to drainage into the ice.

Marginal channels. Marginal channels (Swed. *skvalrännor*) sometimes conveyed melt-waters from the ice itself, now in rock, now in ice, or between ice and rock. More frequently they were lake-to-lake channels joining lakes held up in valleys or embayments; they were often of considerable dimensions, often merely shifting channels clinging to the receding edge. They were

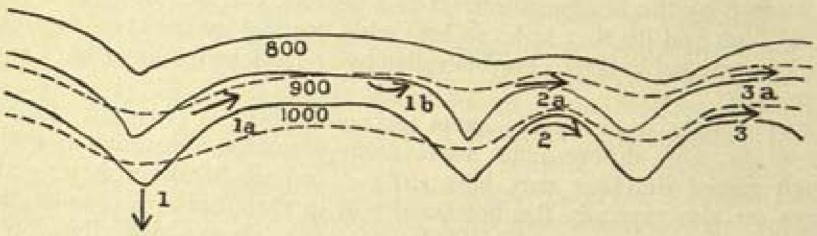


FIG. 86.—Sketch-map to illustrate the different kinds of glacial channels and their relationships. 1, direct overflow; 1b, "in and out" channel; 2, and 3, 1a, 1b, 2a, and 3a in aligned sequence; 2 and 2a, 3 and 3a in parallel sequence.

initiated occasionally as a line of pot-holes flanking a hill,[187] usually and especially on steep hillsides as a shelf or bench without an outer wall.[188] This was provided by an ice-cliff which the glacier's motion renewed as fast as it was melted or eroded: a line of morainic mounds may mark its position.

A shelf of this kind often begins as a cleanly washed hillside,[189] though boulders may bestrew the wall if it is made of till.[190] Further erosion incised the stream in the floor of the shelf and left a low ridge or "marginal footing" along its outer edge.[191] The hollow rapidly evolved into the typical walled valley or marginal channel. Since the ice-face controlled its course, this is usually independent of the present drainage and follows the contours, though occasionally it coincides with rock-structure. Waters diverted around a lobe pressed against a hill cut a crescentic loop or "in-and-out channel". Occasionally, as in the Findhorn area in Scotland,[192] the channel runs exactly along a crest that just separated two adjacent glaciers.

Marginal or submarginal channels are found in series in aligned or parallel sequence[193] (fig. 86). Those in aligned sequence declined in the same direction and functioned simultaneously, their size depending upon the length of the halt of the ice or upon the relief. Their positions fix roughly the average gradient of the edge which was of the order of 1:130 or 40 ft/mile[194] (see p. 42). Channels in parallel sequence were originated as the intermittent or continuous withdrawal uncovered successively lower slopes. The steadily expanding drainage area and lengthening ice-front led to a perceptible increase in the capacity of the channels in aligned sequence and of

the lower ones in parallel sequence. Contemporaneously, the many lesser lakelets at widely differing levels became replaced by an ever-diminishing number of lakes which grew bigger and bigger. The records of water-action are fairly feeble down to the water-partings but left traces in shallow gullies, 1–2 m deep, in places anastomosing at acute angles. Such rills followed in successively lower positions the constantly shifting margin.

Because of their steep gradients, ample loads and frequent if short-lived floods, such channels were probably cut very quickly. This is in keeping with modern observations in Alaska[195] where a gorge 15 m wide and more than 7·5 m deep was cut into rock in 6 years.

Deltas. Overflow valleys at their embouchure into lakes are marked occasionally by cataract or plunge basins, more commonly, as in the modern glacier-lakes that fringe the Cornell Glacier,[196] by deltas of water-worn detritus. These register accurately the level of the lake and the intake of the outlet channel, though the delta's head may be 50 m or more above and its distal end 1 m below this level.[197] The delta is commensurate with the magnitude and duration of the overflow (they are consequently very big in Lake Agassiz[198] and in the glacial Finger Lakes of New York[199]), the material mainly consisting of rocks removed in excavating the valley together with pebbles of the regional drift.

The upper surface is flat or gently sloping and sometimes pitted with kettle-holes, the sites of masses of inert or buried ice. The outer face, with its cusps and re-entrants, is lobate and steep, the steepness being either original or due to waves or shore-currents (sometimes indicated by a sharp deflection of the delta[200]) as the lake fell. The structure is tripartite and the sands and gravels torrential, strongly current-bedded or evenly stratified with clays denoting quieter conditions. The foreset beds may have alternating sands and gravels and clays, possibly seasonal.[201] The contortions, sometimes seen, may be made by ice or by slumping following the drawing off of the water[202]; Swedish studies suggest that deltas of fine-grained sand settled 5%.[203]

Deltas vary from mere spits or ridges deposited on one or other side of the overflow to accumulations which completely clog the glacier-lake. They are very sensitive to emergence arising from the downcutting of the outlet or its replacement by a lower one. Progressive emergence may replace the freely anastomosing streams by definite channels and build a series of new deltas, with their apices beyond the outer margin of the original structure and at points successively farther out. Deltas, as in Lake Agassiz,[204] were seldom constructed where normal land streams poured into a lake unless these were reinforced by melt-waters from local glaciers (pl. XIXA, facing p. 496).

Beaches. Lake-terraces are associated with modern glacier-lakes[205]; a delicate terrace, in places beautifully developed, is traceable around the Merjelen See[206] and another one around Graenalón, Vatnajökull.[207] Those of the smaller or narrower Pleistocene lakes, whose levels fluctuated and fell, are generally faint, rare and impersistent; hard rock formed the shores; the fetch of the waves was small; the controlling ice-border channels were mostly weak and shifting; and later soil movements may have effaced them. Higher inscriptions are particularly elusive. The frozen condition of the lakes during the glacial winters hampered beach development, as did the lake-ice which even in summer covered their bays and branches.

Some of the Swedish nunatak lakes have terraces[208] and Norway[209] has celebrated strandlines in Gudbrandsdal and in the Munkevejene in Trysil. Lapland's beaches supply excellent reindeer pasturage. The vast glacier-lakes of North America, which were suited to wind and wave action, have the most prominent examples; their wave-cut terraces and stretches of rolled shingle extend for hundreds of miles with only an occasional break.

For lake-terraces of good definition, an abundance of beach-forming material and a sufficiently permanent lake-level were necessary. Stability was induced by a favourable topography which presented no alternative outlet during a prolonged and slow retreat,[210] by a hard and resistant rock at the outlet and a long overflow that was only slowly worn down[211] or, as in places in east Greenland, by evaporation in a dry local climate (Heinsheimer, 1954). The sources of the beach were debris rafted by ice-floes and shore-ice; arrested talus washed out by rains and melting snows from the bare ground above the water-line; and detritus eroded by waves[212] which were most active on abrupt slopes or on larger lakes which were parallel with the prevailing wind. The beach was apt to be particularly broad when inflowing streams built out their deltaic fans.

Strandlines notch drumlins, fluvioglacial material and solid rock, especially if this is soft, horizontally bedded or well jointed. Those in rock (Norw. *bergseter*, *inlandseter*[213]) are associated with stacks, caves and undercut cliffs. In till, they are accompanied by boulder accumulations, stranded by floating ice or left behind when waves washed the drift. In North America, they form veritable boulder-pavements a few hundred yards or even half a mile (*c.* 800 m) broad.[214]

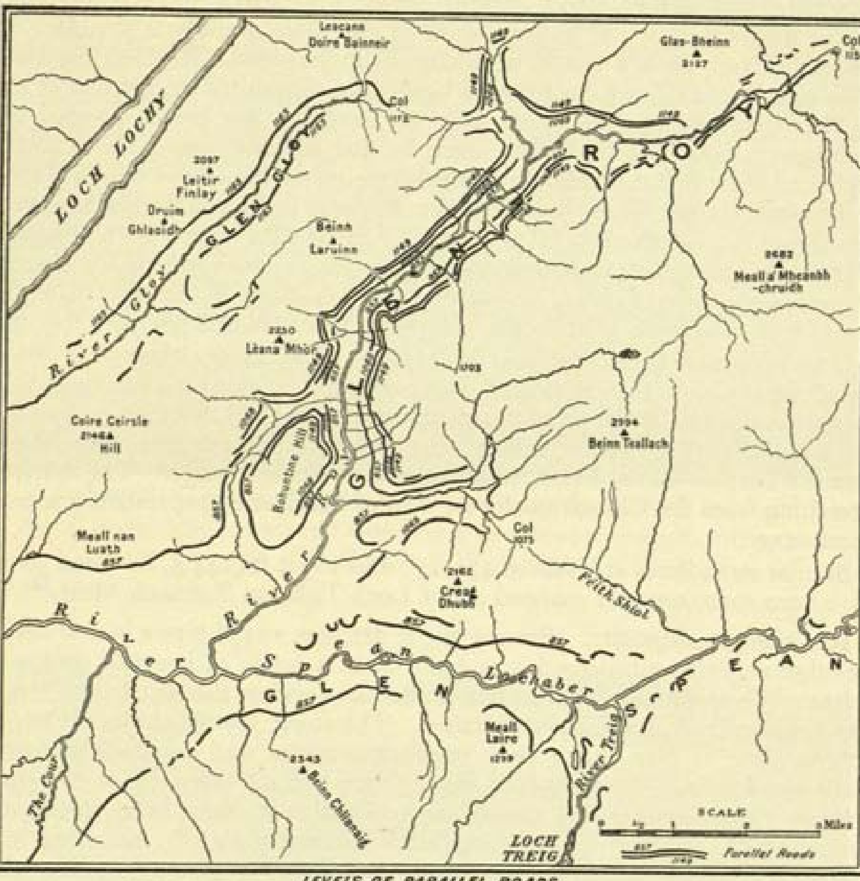
The irregular contours of the shore added spits, hooks and bars[215] from which the direction of the wind and currents may be gauged. Wind-driven floes and expanding lake-ice occasionally created lake-ramparts.[216] Sometimes sand-dunes lie behind the beach of a big lake, e.g. in North America[217] and especially near deltas, though the time was usually too brief to permit their construction.

The terraces, which have been mistaken for those of rivers,[218] but unlike them are or were horizontal, record successive stages in the lowering of the lake by shifting the outlet, less probably by stoping in the bed of the outlet channel[219] or by alternations of dry and moist climates.[220] Tides were negligible even in the vast glacier-lakes of North America.[221] The hollows, 1–2 m deep, which sometimes run down the fronts of large terraces, have been ascribed to the rush of water that resulted when the opening of a new outlet lowered the lake.[222]

Sometimes, as in Denmark and Scandinavia (excepting west Jämtland), terraces and deltas are alike well above the levels of the outlet cols.[223] This "pass anomaly", which may be 10, 20 or even 30 m, has been variously attributed[224] to the occurrence on the col of drift ice, dead ice, inland ice or a local glacier, or, if the discrepancy is less than 6 m, to the depth of the outflowing stream.[225]

Parallel Roads of Lochaber. The classic beaches in extraglacial lakes are the well-known Parallel Roads of Glen Roy, Glen Gloy and Glen Spean, three glens in Lochaber, north-east of Fort William in the Scottish Highlands[226] (fig. 87)—they are represented on the one inch Ordnance Survey map. These roads, used by cattle and man and made of more or less local material, are flat and up to 60 ft (18 m) wide. They slope valleywards at

angles varying between 12° and 30° and sweep round the shoulders of the hills into recesses and tributaries with remarkable horizontality and parallelism. In Glen Roy, where they are magnificently displayed (pl. XXA, p. 497), their heights of 1149 ft, 1065 ft and 856 ft (350, 324·5 and 261 m) correspond with three cols, the one into Glen Spey, as first ascertained by T. D. Lawder and confirmed by Darwin, the second or middle one into Glen Glaster, as D. M. Home[227] first discovered, and the lowest at the head of Loch Laggan, as Lawder pointed out. The highest col had the smallest river, the lowest col the largest outflow.



LEVELS OF PARALLEL ROADS

Glen Gloy					Glen Roy					Glen Spean			
	Highest	Lowest	Average	Col		Highest	Lowest	Average	Col	Highest	Lowest	Average	Col
1	1173	1156	1165	1172	1	1155	1144	1148	1151				
					2	1077	1062	1067	1075				
					3	862	850	855	848	862	850	855	848

FIG. 87.—Map of the Parallel Roads of Lochaber, Scotland. *Mem. Geol. Surv.*, "Grampian Highlands", 1935, p. 75, fig. 22.

The successive interpretations of the beaches reflect the progress of glacial geology. Regarded by imaginative Highlanders as the hunting roads of Fingal and the heroes of his age or as aqueducts for irrigation, they were later deemed to be the flood marks of a violent deluge[228] caused by an earthquake or a subsidence in the Atlantic. Darwin,[229] who had recently returned from South America and its evidence of recent uplift (see p. 1325), considered them

to be ancient sea-beaches. His view (subsequently recanted[230]) received the assent of many[231] but is untenable[232] since the beaches are neither widely distributed nor best developed where most exposed. This fact was noticed as early as 1817 by J. MacCulloch[233] who recognised that the beaches indicated a lake which, as the outcome of a local cause, thrice suddenly subsided. The local cause was the subject of later speculations: these regarded the beaches as river-terraces,[234] as terrace-moraines fringing a valley glacier,[235] or as shores of a lake resting upon ice[236] or imprisoned by a detrital dam 1200 ft (*c.* 360 m) high.[237]

The modern theory of a glacier-lake in the three glens was sketched by Agassiz[238] and worked out in detail by T. F. Jamieson.[239] Glaciers from Glen Arkaig and Glen Eil converged upon Glen Spean (as the content of the Lochaber moraines proves), and aided by ice from Ben Nevis, ponded the drainage in lakes which spilled over by the several cols. The streams deposited their deltas where they entered the lower lakes, as in Glen Turrit at the outflow from Glen Gloy into Glen Roy and in Rough Burn at the overflow from Glen Glaster into Glen Spean. Agassiz's theory has been generally adopted[240] since the beaches and cols coincide in altitude and the former terminate abruptly at points opposite each other towards the mouths of the glens where moraines, skirting the hills, intersect their courses.

J. C. Wilson[241] traced the later history and recognised a number of lower lakes, one at 420 ft (128 m), held by ice spanning the Spean at Tivandrish (the waters overflowed by the col at the south-west corner of Loch Lianachan) and a lower one at 300 ft (90 m) which extended for 5 miles (8 km) from Brackletter to Torr-an-Eas and escaped south-west of Spean Bridge. The moraine stretching from the Cluneston slopes, west of Brackletter, represents a subsequent stage.

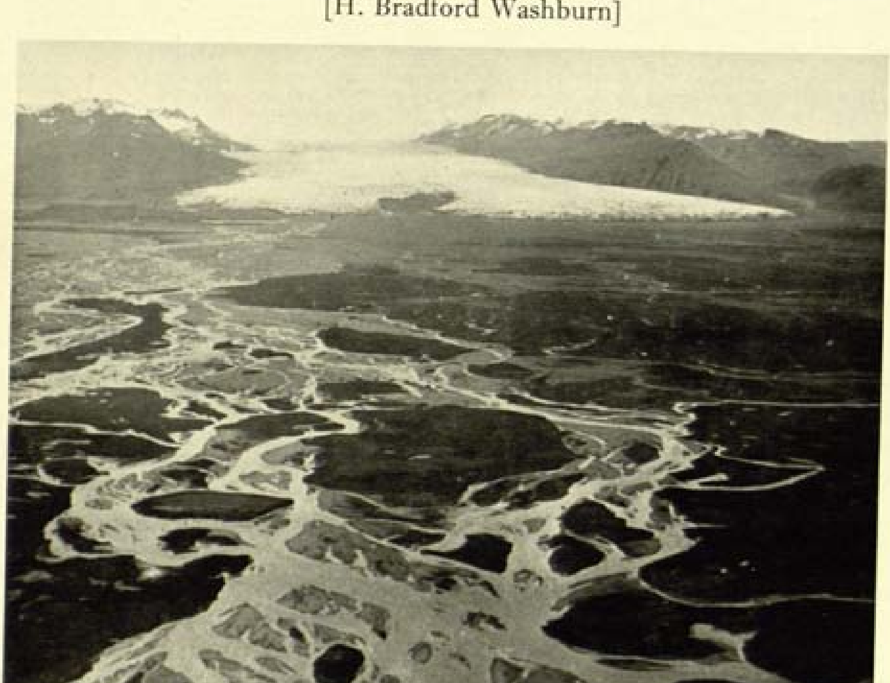
Similar strandlines at levels of 474, 277 and 183·5 ft (144·5, 84·5 and 56 m) have been more recently mapped about Loch Tulla on Rannoch Muir.[242]

Ice-margin deposits. Stream-borne detritus, swept into a lake from an ice-edge partially submerged, was built up in a series of laterally confluent deltas. Their flat tops, 1–2 m below the lake-level at the distal end,[243] are less hummocky than subaerial moraine. The steep, frontal slopes,[244] lobate or crenulate in plan, have deep re-entrant angles and protruding cusps. Such sand-plains,[245] subaqueous fans,[246] ice-contact deltas[247] or marginal terraces (*Randterrassen*[248]), though generally absent, have been frequently described,[249] especially from lateglacial Fennoscandia,[250] including the Salpausselkäs of Finland (see p. 1172); Penck[251] has suggested Salpausselkä as a generic name for these accumulations (pl. XIXB, facing p. 496).

Their material, which has a tripartite structure and becomes increasingly fine lakewards, is finer than that of a subaerial moraine. At the back of the plain are the backset beds which owing to the slow melting of the ice are only one-fortieth or one fiftieth as extensive as the foreset beds. They are accompanied by ice-contact slopes and kettle-holes and contain boulders and coarse gravel, though the ice-edge may be traceable only by a line of plentiful boulders.[252] The mesa-like deltas built out into the lakes may be connected with a feeding esker on the iceward side.

Floor deposits. In striking contrast to the shore features which are seldom well marked, the floor deposits have usually a considerable extent and thickness; they may completely fill the water-bodies.[253] As seen in modern

A. Woodworth Glacier, Alaska, with two recently formed osar and (*on right*) seemingly ice-scored ridges in a frozen apron of ground moraine
[H. Bradford Washburn]



B. Outwash with braided streams, Flaajökull, Vatnajökull, Iceland
[Royal Geographical Society]

A. Newton Dale overflow valley, Cleveland Hills, looking upstream [Geol. Surv. Gt. Britain : Crown copyright]

B. Outlet channel of Lake Agassiz on boundary between Minnesota and South Dakota, north of Lake Traverse [R. J. Lougee]

(*facing p.* 465)

and Pleistocene glacier-lakes,[254] they consist of silts and clays (Swed. *issjölera*; Dan. *issøler*, *Fladbakkeler*) which represent the impalpable milk, like that which to-day discolours glacial lakes in the Yukon district over a distance of 30 miles (*c.* 50 km),[255] and the finer materials transported from the ice-free slopes above the lakes or removed in cutting the overflow valleys. These muds are purest in the centre of the lakes where their countless laminae, the book-leaf clays (Ger. *Bänderton*), are greatly persistent, regular and exceedingly thin, up to 200 laminae occupying 1 in. (*c* 25 mm). Their surface is shiny when cut vertically with a knife and displays alternate ribands of lighter and darker hue. The fineness varied with the seasonal or periodic rise and fall in the volume of the streams. Laterally, they pass imperceptibly into the distal parts of the beaches and ice-margin deposits and become increasingly sandy; sands and clays may indeed take their place on the floor. Changes of sedimentation were apt to occur in lakes which discharged intermittently through crevasses and had shallow and deep episodes.

The ultimate destination of much of the mud was the sea; thus the German *Urstromtäler* carried much silt into the North Sea.[256]

Ice-rafted erratics. In deep lakes, like those the Greenland Eskimos call *iluliatik*, i.e. "provided with icebergs"[257], occasional boulders are rafted out by small bergs from the ice-face and by shore-ice,[258] as in Merjelen See, the lakes held up by the Cornell Glacier, or in numerous Pleistocene glacier-lakes[259] where stones were sometimes floated beyond the limits of glaciation. The "haystacks" or *saxums* scattered over the floor of Lake Leverett (see p. 479) were ice-rafted boulders.[260] While the erratics may be irregularly scattered over the floor or be embedded in its sediments, they are apt to concentrate at the entrance to the outlet[261]; their maximum altitude fixes roughly the lake-level.[262] In some sections of lake-warps the erratics diminish upwards, suggesting that floating ice became less common as the ice shrank back.

Drift-ice, by grounding, distorted the laminae and by melting made small pits ("iceberg hollows"[263]). It sometimes left its load as "nests" of erratics[264] or as till-like deposits lying upon or between the lake-sediments.[265]

Oscillatory margins. A positive oscillation during a general recessive phase may modify the various features associated with marginal lakes. The ice may override and block lower channels or deepen older ones and make new ox bows and higher channels, as in modern Alaska.[266] It may unite and bifurcate channels in an apparently inconsequent manner, or bury them, as in present-day Alaska[267] or Pleistocene Scotland.[268] It may deposit till or moraines upon them, especially near their mouths, or upon beaches or lake-muds[269] and incorporate or contort some of these.[270] The rising waters may cause a higher beach to succeed a lower one, as in "Lake Passaic",[271] lay down lake-silts on earlier sand and gravel beaches,[272] or sandwich a thick layer of sand and gravel between two horizons of fine lacustrine silt, as in "Lake Pickering",[273] Yorkshire.

Crumpled, veined and faulted lake-deposits, though often denoting an advance, may also result from grounding bergs, from melting buried ice, from flow-distortion of bedded mobile mud set in motion intermittently towards temporarily opened crevasses, or from landslides in deeply entrenched clays which sometimes injected sands into adjacent clays.[274] In this connexion it is necessary to bear in mind that all drift was originally saturated with water and that mud-flows and slides before drying must have been common.

Glacier oscillations were generally climatic but sometimes, as has been suggested for certain advances[275] in Baffin Land and Fennoscandia, were governed by a lowering of a lake: by putting an end to calving, the ice advanced over lake muds in the calving area.

(b) *Regional Distribution of Pleistocene Glacier-lakes*

Early phases. Melt-waters were probably active during the various advances and recessions of the ice-sheets preceding the last and were pent up in glacier-lakes[276] which rose during the advancing hemicycle in ascending order as the ice sealed lower outlets. The evidence of these events has been mostly obliterated within the terrain covered by the later ice: the beaches were a ready prey to ice-erosion, overflow channels were blocked with drift and floor-silts were overridden and incorporated in later boulder-clays. Outside these areas, it has been concealed under later outwash.

Buried overflow channels have been discovered in the Finger Lakes region[277] and in the Ohio and Erie basins[278] and lake-silts beneath drifts in several North American localities.[279] Wave-cut terraces of ice-dammed lakes underlie Wisconsin drift,[280] and spillways, vigorously ice-eroded and veneered with till or choked with various kinds of drift, are found in Washington and New England.[281] Early glacial col gullies may occur in Torne Träsk and Lapland.[282] Lakes were impounded by Illinoian ice in Pennsylvania and Illinois on the older drift[283] and for a short time in Iowa (Iowa and Cedar valleys) on the Illinoian drift[284] by the displacement of the Mississippi ("Lake Calvin"). A vast proglacial lake existed in Kentucky and West Virginia in pre-Illinoian time and lake-sediments of Nebraskan age are known from Ohio and West Virginia.[285] Anomalies in the through valleys of south-east Ohio indicate multiple cycles of Pleistocene cut and fill.[286]

A lake-history as complex as that of lateglacial time marked the advance and recession of the pre-Wisconsin ice in the region of the Great Lakes (see p. 1351). The outlets of the retreat were not necessarily those of the lake occupying the same basin during the advance: ice may have lowered the rim or may have raised it by deposition.[287]

An early glacio-lacustrine phase, with marginal channels, existed on the great plains of Canada[288]—Lake Agassiz had an interglacial predecessor[289]—and in north Germany and the country south of the Baltic,[290] viz. in the Weser area during the Elster glaciation and west of this during the Saale glaciation; boulders were rafted out into extraglacial lakes during the advance.[291] The Lauenburg Clay has been attributed to a glacier-lake[292] and the Dnieper and Don lobes in Russia found a system of marginal stream-channels belonging to the third glaciation.[293] The Alpine glaciation has yielded other instances of early glacier-lakes.[294]

Yet this fragmentary evidence does no more than hint at such constrained drainage during earlier phases. It may indeed be that the conditions during the advance and during the retreat differed essentially, the climate being persistently dry until after the withdrawal began.[295] Thus the sands and gravels interbedded with the boulder-clays of East Anglia are associated with the underlying rather than with the overlying boulder-clays. However that may be, the events now to be reviewed belong in general to the closing phases of the Glacial period (see chs. XLII, XLIII).

Circumbaltic region. The glacier-lakes which margined the Scandi-

navian ice in a gradually contracting ring drained westwards over Poland and Germany to the North Sea; southwards (with their erratics) into Bohemia,[296] the Danube and Black Sea; by the upper Vistula into the Dnieper[297] which carried the waters from 1000 km of ice-edge and 600,000 sq. km of ice-surface; by the other rivers of central Russia into the Caspian Sea[298]; and northwards across Russia into the White Sea and Arctic Ocean. The drainage, therefore, was centrifugal in the central strip and marginal or peripheral in the west and north.

Maximum glaciation saw glacier-lakes where the Scandinavian ice impinged upon the northern flanks of the German Mittelgebirge. The most westerly, which had floor deposits, rafted erratics and spillways, lay in the Meuse and Rhine[299] while the ice stood at Krefeld; it only disappeared when the ice fell back east of the Ijssel. The ice crossed the Rhine, pressed against the Rhine-Main terrace (see p. 1043), and floated its Scandinavian erratics southwards.[300] Big erratics, patches of till and disturbed beds are found south-west of the river[301]; a frontal moraine (= Rehburg stage) runs between Nijmegen and Krefeld[302] (see p. 711); and the terraces in the Rhine and Meuse are interconnected through several depressions[303] (*Oerstroomdalen*) as by Gennep and the River Niers (= Middle Terrace). Similar lakes drowned the valleys of the northern slopes of the German Mittelgebirge[304] (their clays rest on the "preglacial" terraces), and the Carpathians of Galicia[305]: the lake in the Mulde was 30 km long and that in the Elbe extended beyond Prague.[306] The Saale ice ponded the Weser and diverted the drainage through the Porta-Osnabruck valley.[307] These lakes were impounded by both the Elster and the Saale ice-sheets.[308]

Later lakes are bound up with the north German *Urstromtäler* (fig. 88) to which Hoffmann directed attention in the first quarter of the last century. L. v. Buch and H. Girard[309] added to the knowledge of the valleys, Girard recognising three of them. These big, peripheral valleys are cut into the *Geest* (see p. 442) and were occupied during the recession by powerful rivers, swollen with the normal drainage from the ice-free country on the south and with melt-waters from the melting of many cubic kilometres of ice on the north and the precipitation from 800,000 sq. km (see below). The most important are: (1) The Breslau-Magdeburg valley (of K. Keilhack) which extends westwards from the Pilica along the Oder south of the Fläming (Warthe) Line (see p. 943) and continues eastwards with corresponding moraines into the area of Czenstochau[310]—it included the Bober, Queis, Neisse, Spree, Black Elster, Elbe, Ohre, Aller and lower Weser; (2) the Glogau-Baruth valley (of E. Geinitz) which skirts the northern edge of the Fläming, south of the Brandenburg line, and continues eastwards into Poland[311] and westwards into the Warsaw-Berlin valley[312]—it carried off at the Brandenburg stage *c.* 30,000 cu. m/sec (size of the River Niger and about 40 times the volume of the present Elbe) and in the warmest month *c.* 41,000 cu. m/sec (size of Ganges-Brahmaputra), derived from the melting ice-sheet to the north and the drainage from the rivers to the south from the Elbe to the Vistula and the Bug and Memel and Minsk in the east, and during the retreat probably 10 times this volume; (3) the Warsaw-Berlin valley (of G. Berendt) which runs from the Bug and Narew and along the Vistula to the Oder, Spree and Elbe in front of the Frankfurt stage—it continued into the drainage system east of the Baltic[313]; (4) The Thorn-Eberswald valley, traceable from the Vistula at Thorn along the Netze and Warthe to the Oder and by the southern

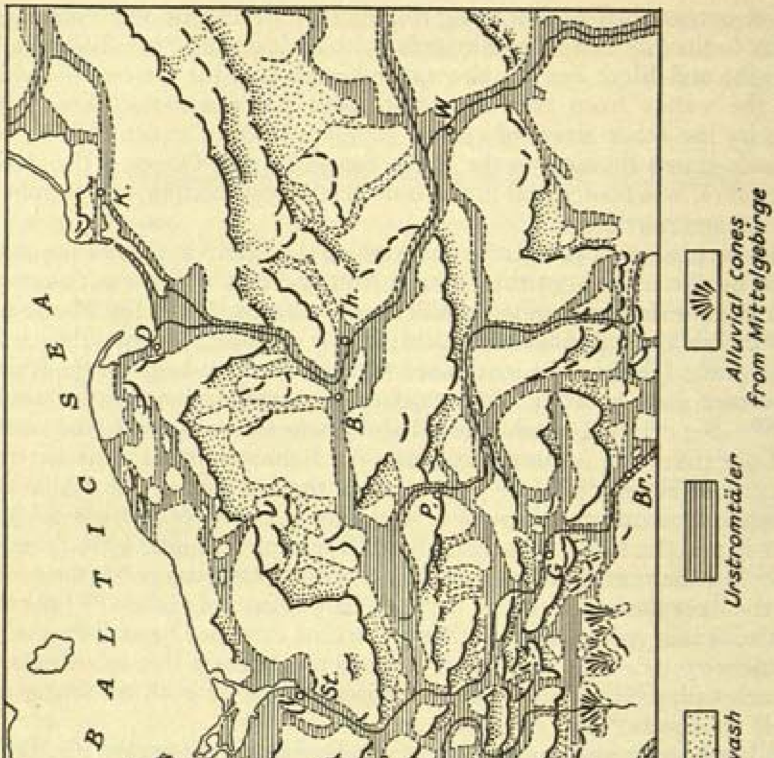


FIG. 88.—Map of the North German *Urstromtäler* in their relationship to the main ice-stages. P. Woldstedt, *1822*, p. 36, fig. 5.

foot of the Mecklenburg *Seenplatte* and the Eberswalde Pforte to the Elbe (thought to be a contemporary of the Pomeranian stage,[314] earlier than the Baltic end-moraine,[315] two different valley systems[316] or not an *Urstromtal*[317]); (5) the Pomeranian valley,[318] north of the Baltic Ridge, interpreted by L. Finckh and O. Schneider[319] as a series of valleys eroded in succession as the ice shrank northwards. The last of the lakes lay in the Haffs,[320] e.g. in the Kiel and Lübeck area,[321] in the Stettin Haff,[322] in Danzig Bay,[323] the

lower Vistula,[324] at Königsberg[325] (with lake-warps or *Deckton*), and the Bay of Riga.[326] Overflow streams may have cut the Danish straits (see p. 1290).

The major *Urstromtäler* converged upon the Elbe since the ice fell back more rapidly from east to west.[327] It is, however, uncertain whether the Elbe drainage was also deflected westwards along the Bode, Ilse and Oker valleys into the Weser; for while some believe this,[328] others maintain that the valleys are at the wrong height or lack the characteristic shape.[329] A Hunte-Leda *Urstromtal* between the lower Weser and the lower Ems has been conjectured.[330]

These waters escaped into a lake in the southern part of the North Sea (see p. 1230) and entered the ocean south of Land's End by the "English Channel River" (see p. 1231) whose volume fluctuated with the season and was of gigantic proportions: it carried the precipitation from the ice-free country in east England and west Europe (e.g. Thames, Meuse, Rhine, Elbe, Oder, Vistula) and melt-waters from thousands of kilometres of ice-edge.

This interpretation of the *Urstromtäler* has been widely accepted[331]; they skirt moraines, have discordant distributaries and continuous terraces, and possess outwash cones built out by streams from the north and deltas laid down by rivers from the south. Oscillations, masses of dead ice, cones and deltas deposited by lateral streams and considerable dunes explain why the valleys do not fall continuously from east to west[332] (see p. 506). Subsequent earth-movements may also have played a part.[333]

Yet this view, though almost certainly true for the southermost (Glogau-Baruth) valley, is by no means unanimously held. The valleys, it is said, are preglacial[334] or interglacial[335] and bear no relation to moraines or melt-waters[336] or to streams coursing simultaneously along their whole length.[337] They were excavated by an extensive system of ramifying subglacial streams[338] or, if used by glacial melt-waters, coincide with pre-existing tectonic valleys[339] (this is stoutly contested[340]) as bores and their parallelism with Hercynian folds suggest.

In the opinion of most German geologists, each valley is older than the one immediately to the north—the southernmost one is the oldest—and marks a definite recessional stage when the ice halted along the whole line of the ridge to the north. But this view admittedly encounters difficulties. There is, in the first place, the problem of the *Durchbruchtäler* or how exactly the drainage broke through northwards, e.g. the break of the Oder through the Brandenburg stage at Neusalz, the Frankfurt stage at Frankfurt and the Pomeranian stage near Posen, and of the Vistula through the Frankfurt stage near Plock and the Pomeranian stage near Neuenburg. It may have been accomplished by tilting, due to the Littorina subsidence[341] or Baltic subsidence[342]; by streams which worked back southwards or sawed their way through the watersheds from the ice-edge[343] or flowed along this as the ice fell back, beginning possibly subglacially[344]; or by subglacial streams which drained northwards along pre-existing valleys kept open as tunnels by warmer waters from the south[345] (the temperature of the waters and their tendency to choke their beds with detritus render this unlikely). Cones of debris suggest that the streams flowed outwards from the ice.[346] Finally, the *Urstromtäler* are regarded as either younger[347] or older[348] than the *Rinnenseen*.

Many facts requiring sifting and co-ordination give an indistinct picture of the glacially constrained drainage of Russia. A. G. Högbom,[349] for example, suggested that several ill-defined glacier-lakes existed in central Russia during

maximum glaciation between the Ural ice and the Scandinavian ice in the region of the Wischera River; the big valleys carrying canals, e.g. between the Volga and Ladoga, the Volga and Onega, and the Dwina and Koma, are probably spillways. There were later lakes in the Pinega, Dwina and Vistula, the latter draining westwards into the north German system, the others southwards into the Caspian Sea. Lakes with banded clays lay in the upper and lower Vistula, as about Sandomierz, Pulawy and Warsaw, and in the Bug,[350] and were connected by a system of *Urstromtäler* (Pol. *Pradoliny*) in west Poland.[351] Another lake occupied the upper Dnieper[352] and spillways fell northwards into a lake about Nijni Novgorod.[353] Other lakes and channels have been found in the Pripet Marshes,[354] in the Warthe valley,[355] in Lithuania,[356] east of the Baltic[357] and about Grodno.[358] North-west Russia has its quota,[359] including a Ladoga-Onega Lake.[360] The Dwina was ponded and diverted through a Linega glacier-lake into the Arctic Ocean. Valleys opening westwards in the Kola Peninsula have lake-terraces up to 400 m.[361]

Extraglacial lakes in east Jutland and Denmark drained northwards along the edge of the ice[362]; one of the most important stood near Stenstrup, Fyn.[363]

Lakes on the wings. In Holland,[364] the falling back of the ice allowed the rivers to move northwestwards in stages: the Rhine followed first the Niers valley, then broke through by Elten and Nijmegen (after retreat behind the Veluwe) and finally divided into the Ijssel and Lower Rhine, a bifurcation that was changed only in historic time.

In Asia the northern ice, as T. Belt[365] anticipated, imprisoned the northerly flowing rivers, though how they overflowed is at present unknown. G. J. Tanfiljew (1902) and J. A. Moltshanoff (1926) inferred the existence of a vast glacier-lake stretching from Lake Baikal to the Caspian Sea and from the Ural Mountains to the Yenisei and Ob. Draining into the Aralo-Caspian Sea (see p. 1131) by the Turgai depression, it may have conveyed the relict fauna into Lake Baikal (see p. 1417).

Scandinavia. The strandlines of the Scandinavian ice-lakes were noticed by C. Linné, R. Chambers, B. M. Keilhau and many others; later geologists[366] related the beaches to the passes and spillways. The lakes have now been mapped in Finland[367] (including the Carelian Peninsula), north Norway,[368] Sareks region,[369] Jämtland[370] and south Sweden[371]: B. E. Halden[372] has given a comprehensive account of them. The general distribution of the glacier-lakes in Baltoscandia is given in fig. 268, p. 1289. Three stages, not everywhere strictly contemporaneous, may be distinguished[373]: (1) nunatak lakes,[374] held up along the sides (especially south) of the nunataks; (2) lakes, up to 110 km long,[375] impounded by eccentric ice (see p. 668) in the big valleys east of the main Scandinavian divide; and (3) drainage, now marginal, now submarginal, along the slopes of the Swedish valleys into the Gulf of Bothnia (Bothnian stage[376]) or with an *Aftappning* under the ice, such subglacial streams being partly responsible for the present valleys,[377] for some of the osar east of the iceshed,[378] and for the giant varves in Jämtland.[379] The direct overflows of stage 2 crossed the cols into Norway. They are now indicated by passes, washed free of moraines, e.g. Bardodal, north-west end of Torne Träsk, Hundal on the Ofoten railway, and Brudslöjan, west of Storlien, and by deep, narrow canyons, with dry or "dead" falls (*døda fall*) on the Norwegian side, their material filling the rock-basins and building deltas at

the level of the then sea[380] (fluvioglacial and early glacial erosion have been held responsible for the cols[381]).

Lakes also existed in the south Swedish uplands[382] and east of the watershed in south Norway[383]; the ice closed the outlets to Oslofjord and diverted the drainage into Guldal, Drivdal and Romsdal.

Alpine glaciation. Constrained glacier-drainage was generally rare in the region of the Alpine glaciation for reasons already noticed (see p. 456). The best evidence, often in the form of channels, has been obtained about Bodensee,[384] in the Danube basin between Sigmaringen and Biedlingen,[385] in the Inn, Ill and Etsch[386]—some of these led J. Brunhes to his hypothesis of erosion of U-valleys by lateral streams (see p. 330)—the Bavarian Alps,[387] west Switzerland,[388] the flanks of the Jura Mountains[389] and near Lugano.[390]

Rest of the Old World. Although studies of the glacier-lakes and their histories figured largely in American glacial literature of the later part of the 19th century, the lakes of the British Isles, with the exception of the classic lakes of Lochaber (see p. 462) and an occasional notice, e.g. for the Lake District[391] and the valleys of the Aire[392] and Clyde,[393] were neglected until Kendall's investigations in the Cleveland Hills[394] and Cheviot Hills.[395] These were the starting point of many British researches applying his methods (see ch. XLIII). Glacier-lakes of Pleistocene age elsewhere in the Old World have been detected in Iceland,[396] the Himalayas[397] and in New Zealand[398] where they were up to 17 miles (*c.* 28 km) long and have varves registering 13,680 years.

North America. Beaches, the most striking feature of the North American glacier-lakes, attracted geological attention in the 1840s[399]—the associated sand-bars, off-shore bars, clays and sands were noticed a decade later.[400] The early work on the beaches revealed two schools of thought. The first, represented by Lyell[401] and subsequently by J. W. Spencer[402] and R. Chambers,[403] regarded them as marine. The second, championed by J. S. Newberry,[404] attributed them to ice-dammed lakes, a view afterwards strongly emphasised[405] and proved by the discovery of the outlets and freshwater shells (see p. 458). A third hypothesis, linking the lakes with land barriers erected by differential tilting, never won more than occasional approval[406] (pl. XXb, facing p. 497).

It is now agreed that a series of gigantic freshwater lakes was situated between the southern margin of the Laurentide ice and the St. Lawrence-Mississippi divide over a front of more than 1000 miles (*c.* 1600 km). The beaches are related to outlet channels (first demonstrated by Gilbert[407] in 1871), both being associated with moraines delineating the ice-front as Leverett[408] established for Lake Warren in Ohio and New York and Taylor[409] for Michigan. In general, the beaches end abruptly or gradually on the north and east at the ice-margin of the time, but extend farther and farther north as the strandlines become lower with the recession of the barrier. The exceptional burial of strandlines by lake-floor deposits proves temporary re-expansions of the ice and the closing of earlier and lower outlets.

Systematic exploration dates from 1887 when Spencer[410] began his investigations. Many workers have gathered the material which, summarised by Gilbert[411] in 1890, enables us to appreciate the extent and succession of the lakes. They lay in five regions: (1) the Great Lakes region; (2) Lake Agassiz

in the Red River basin; (3) the Hudson Bay drainage in north Ontario; (4) the western Cordillera; and (5) about the southern ice-centres.

Great Lakes region. Numerous lakes, bewilderingly complex in their history and nomenclature and initiated in the Cary substage of the Wisconsin

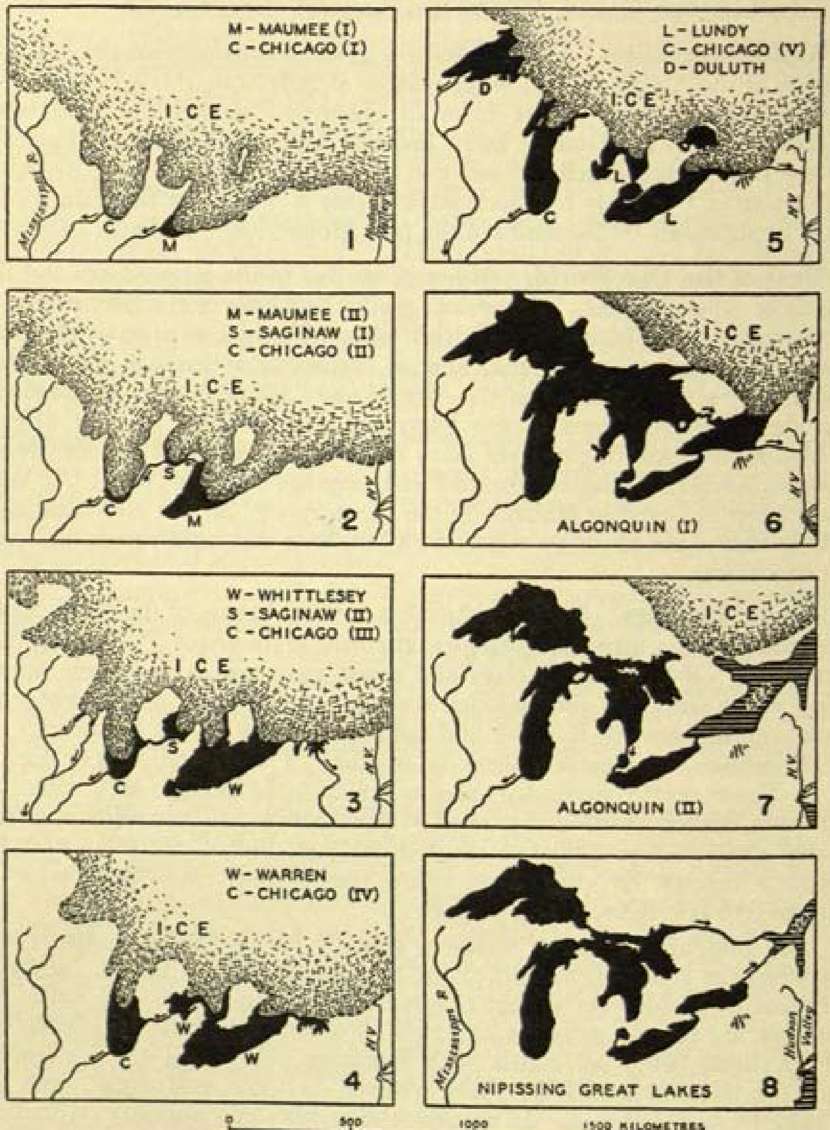


FIG. 89.—Eight stages in the history of the glacier-lakes of the Laurentian basin. R. A. Daly, *338*, p. 98, fig. 54.

glaciation (see p.970), existed in the region of the Great Lakes (fig. 89). Their history, which was controlled by the ice-recession, by crustal warping and by erosion of the outlet channels, has become more and more complex with increasing knowledge, the discovery of numerous short-lived phases, and the recognition of recurrent readvances,[412] such as that suggested by cores from

the deeper parts of Lake Michigan which prove a low-water stage believed to be post-Algonquin and pre-Nipissing. Only the principal episodes can here be mentioned (cf. table facing p. 478).

Extra-limital lakes were held up at the maximum of the last glaciation, e.g. in the Driftless Area of Minnesota,[413] in Wisconsin[414] (Lake Wisconsin covered nearly 2000 sq. miles; 5000 sq. km) and in south-east Ohio and adjacent parts of Pennsylvania, West Virginia and Kentucky where abandoned valley remnants attest extensive drainage changes about the periphery of the continental glaciation.[415] During the earlier phase of the withdrawal, the Laurentide ice shrank back up the slopes leading from the Mississippi and within the limits of the Great Lakes basin, the ponded waters escaping in numerous places across the rims. As the ice receded, the waters grew and united along the retreating edge. Thus crescentic lakes occupied the extreme ends of three of the great basins[416]; they were "Lake Duluth" at the western extremity of Lake Superior (Lake Duluth had smaller precursors in the St. Louis and other valleys along the ice-edge[417]) which drained first by the St. Croix river into the Mississippi—erosion of this outlet lowered the lake through several levels marked by separate strandlines—and later into the Michigan basin (Lake Chicago) by channels skirting the eastern flanks of the Huron Mountains[418]; "Lake Chicago", at the southern end of Lake Michigan, which had three distinct strandlines and drained by the "Chicago Outlet",[419] a valley in drift, *c.* 21 m deep, 3–5 miles (*c.* 5–8 km) broad, and 300 miles (480 km) long, that carried the overflow across the col near Chicago into the Des Plaines, Illinois and Mississippi rivers—Lake Chicago originated as a string of lakelets at the edge of the ice when the inner slope of the Valparaiso moraine began to be laid bare and ceased when the uncovering of the peninsula between the Huron and Michigan basins allowed the lake to merge into Lake Algonquin (see below); and "Lake Maumee",[420] at the western end of Lake Erie, which has two strandlines and carried the waters from the Erie and Huron lobes from Pennsylvania to the base of the Thumb of Michigan; it discharged over the divide at Fort Wayne into the Wabash River across Indiana and later (after a withdrawal of the ice down the Thumb) by the Imlay Outlet and the "Grand River" (J. H. Bretz, 1951) into "Lake Chicago". "Lake Passaic"[421] in New Jersey and the lakes in Connecticut[422] and Massachusetts[423] were probably their contemporaries.

The Grand River and later channels farther north in the Thumb of Michigan near Ubly ceased to function as spillways during the drop to Lake Wayne levels and when the Two Creeks Forest Bed (see p. 972) was formed (= Bowmanville stage) and before the Calumet beach in Lake Chicago was built. The readvance that followed and closed the Lake Mayne outlet caused the glacial lakes to return to their former levels and the Chicago Outlet to come into use once more.

"Lake Whittlesey",[424] a later stage of Lake Maumee but separated from it by the low-level "Lake Arkona"[425] (that marked the end of the Cary substage), discharged by the Ubly channel into a "Lake Saginaw" in Saginaw Bay. The two lakes afterwards coalesced and drained into Lake Chicago across the Thumb of Michigan by the Grand River and successive outlets.[426] Lake Whittlesey, which expanded northwards and westwards with the recession, became "Lake Warren"[427] after a low-level "Lake Wayne" stage,[428] whose beach is much smoothed and reduced by later submergence and whose outlet was by a plexus of channels south of Syracuse to the Mohawk and

Hudson. Lake Warren, after mingling with Lake Michigan and uncovering the watershed between it and the lake in the western end of Lake Superior ("Western Lake Superior"[429]), blended into one vast sheet which discharged through the Grand River outlet into Lake Chicago. Leverett's maps[430] show the position and extent of lakes Maumee, Whittlesey and Warren.

Contemporaries of Lake Whittlesey were probably the glacier-lakes, up to 65 miles (104 km) long, in the Adirondacks,[431] and those in the Finger Lakes region[432] which drained through magnificent spillways southwards into the Susquehanna, Mohawk and Hudson, westwards into the Mississippi drainage, and north of the Adirondacks north-eastwards into Lake Champlain, the Hudson and St. Lawrence. A complicated series of changes was passed through before a dozen or more of these separate lakes, each confined to its own valley, united to make "Lake Newberry" at *c.* 900 ft (*c.* 275 m), which overflowed southwards from the head of Seneca Lake valley to the Susquehanna, then "Lake Hall", which drained westwards to Lake Warren in the Erie basin, then "Lake Vanuxem", which was tributary past Syracuse to the Mohawk.[433]

Further recession resolved Lake Warren into three lakes: "Lake Algonquin"[434] which drowned much of the Superior, Michigan and Huron basins; "Lake Erie" in the Erie basin; and "Lake Iroquois" in the Ontario basin. Lake Algonquin, up to *c.* 425 m deep, was the biggest of all the Laurentian glacier-lakes: it covered the three upper lakes and much land on the north and east. Its well-defined beach, traced by Spencer and Taylor[435]—other beaches occur at lower levels[436]—is like those of the other big lakes well known except on the north where the waters met the waning ice and the country is largely forested and ill-equipped with roads—in a general way, the Canadian Pacific Railway, situated upon its sands and gravels (see p. 501), marks its limit on that side. Its overflow was at its earliest stage (Algonquin I, when the lake covered southern Lake Huron) by Port Huron and the Chicago Outlet,[437] and later (Algonquin II) into Lake Iroquois by the "Algonquin River"[438] or "Kirkfield Outlet", a deep and wide channel across Ontario (this is cut to below the present level of Lake Ontario at Trenton[439]). A differential isostatic uplift of the northern part of the basin (see ch. XLV) or a readvance of the ice raised the lake's level so that it once more spilled over both at the St. Clair River and Chicago outlets[440] (Algonquin III); for a time there was a two-outlet phase when the "Algonquin beach" or the highest Algonquin beach was formed and the land was still canted to the south. This stage terminated not by tilting but by lowering as the retreating ice uncovered a succession of lower outlets[441] between Georgian Bay and Lake Ontario basin (Algonquin IV). The lower water-planes (Wyebridge, Penetang, Cedar Point and Payette) correspond to temporary halts in a declining lake.

Lake Iroquois, the best known and the first of the lakes to be defined,[442] probably began about the same time as Lake Algonquin[443] but failed to survive this lake.[444] Over 300 m deep, it existed for about 8000 years[445] (its shore, which is missing over about 70 miles (*c.* 110 km) on the north-east where the ice-barrier stood,[446] is about as mature as that of the present Lake Ontario) and sent its surplus waters by Rome into the Mohawk and Hudson[447]—thereby lowering the gorge at Little Falls and building up the broad delta in the Hudson Valley[448]—and later by the North Bay Outlet. It subsequently sank *c.* 88 m into the short-lived "Lake Frontenac"[449] (marked by varves and beaches) and later into "Admiralty Lake" (see below).

The ice in the Ontario basin, in melting back from the Niagara escarpment, formed during one of its stages a "Lake Lundy" (Lakes "Dana" and "Dawson" in New York) which was succeeded by Lake Iroquois when the Niagara cuesta was left above the waters. Thus Lake Erie became a separate body, discharging by the Niagara River, and Niagara Falls came into being (see p. 1518).

Crustal warping and a withdrawal of the ice from the lower St. Lawrence opened a direct outlet to the east and an escape at North Bay, Ontario, by the Mattawa valley.[450] The Nipissing Lakes,[451] outlined by the extremely sharp Nipissing beach, now occupied the basins of Superior, Michigan and Huron; the waters stood at the same level and covered the present rapids of Sault Ste Marie to a depth of 15 m (over a width of 8 miles or *c.* 13 km) and the Strait of Mackinac to about the same depth.[452]

The Nipissing Lakes were "postglacial" and independent of ice-barriers. They were slightly bigger than the present lakes since their beach (which is unconformable with the earlier beaches[453]) is usually less than 1 mile (*c.* 1·5 km) from the water's edge of to-day. They were fed in the early stages by waters from the receding ice, later by overflows from "Lake Ojibway" (see below). During the earlier phases, the three upper lakes drained by the "North Bay Outlet" along the Mattawa River and over the site of Ottawa into the Ottawa Sea and St. Lawrence valley (see p. 1309). For most of the time there was a two-outlet phase when the drainage also escaped by the St. Clair-Detroit River outlet to Lake Erie. At the close of the period, when progressive uplift in the north diverted the discharge from the North Bay outlet to Port Huron and via Niagara into Lake Ontario, the three great lakes much resembled in outline their modern successors. When the ice withdrew from the Thousand Isles, the Ontario basin was occupied not by the sea (as formerly thought) but by a freshwater Admiralty Lake[454] which received the effluent of Lake Algonquin: the sea then stood much lower.

J. H. Bretz[455] has suggested that poor lake-beaches were formed in periods of general down-cutting of outlets due to retreats when much water was released and cubic miles of ice melted, and good beaches in periods of advance when the volume of water decreased and an armour of boulders shod the outlets.

Lake Agassiz. The most extensive of all the North American extra-glacial lakes was "Lake Agassiz", a close correlative of Lake Algonquin[456]; its length was 600–700 miles (*c.* 960–1120 km), its breadth 200–250 miles (*c.* 320–400 km), and its area 110,000 sq. miles (*c.* 285,000 sq. km)[457] or that of the existing Great Lakes combined. About 200 m deep above the level of Lake Winnipeg, and of a duration estimated at 1000 or 10,000 to 15,000 years,[458] it occupied the basin of the Red River and the upper Assiniboine, covering much of Minnesota, north-east Dakota, Saskatchewan, Manitoba and Ontario (fig. 90) with its rich alluvial deposits, the wheatlands of to-day—Lake Winnipeg and some smaller lakes, e.g. lakes Cedar, Winnipegosis, Manitoba and Lake of the Woods, nestle in hollows on its floor. The lake persisted until the ice-front which then trended about north–south had withdrawn east of the middle course of the Nelson River.

A lake on this site was postulated as early as 1823 by W. H. Keating and afterwards by several geologists, but it remained to W. H. Winchell[459] to show that it was ice-dammed. For our present knowledge we are indebted

to Upham[460] and his successors,[461] including the Canadian geologists who have traced its northern extent.

When the ice shrank from the parting between the Red River and the Mississippi, several small lakes, e.g. lakes Minnesota, Saskatchewan, Souris and Dakota, came into existence[462]—independent lakes, with melt-water channels, lay in Alberta south of Edmonton and drained across the continental divide between the Arctic and Gulf drainage into the Missouri.[463]

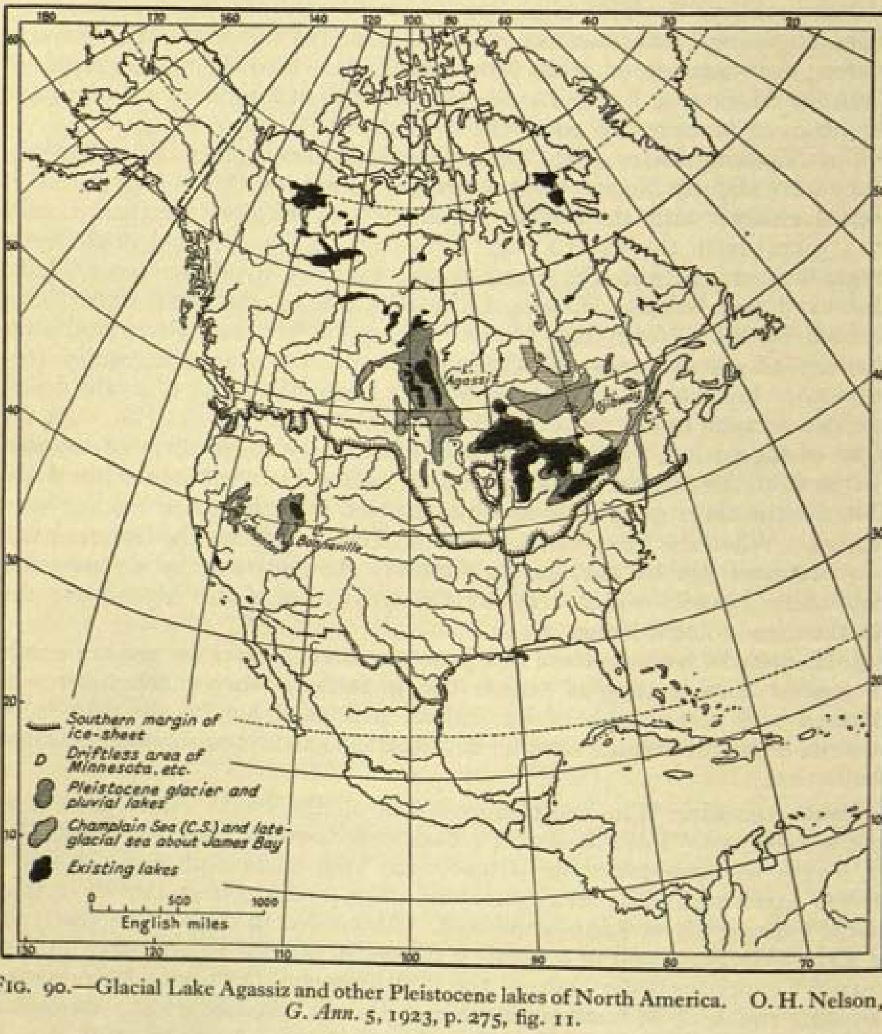


Fig. 90.—Glacial Lake Agassiz and other Pleistocene lakes of North America. O. H. Nelson, *G. Ann.* 5, 1923, p. 275, fig. 11.

With continued retreat of the Red River lobe of the Mankato ice and probably about the time of Lake Duluth, these lakes joined into a larger sheet, G. K. Warren's "Lake Agassiz", retained between the high ground on the south and west (the Manitoba escarpment) and the Keewatin and Labradorean ice-sheets united in the lower course of the Nelson and Churchill rivers.[464]

The lake, possibly augmented by streams from glacier-lakes in the Peace and Athabasca rivers,[465] drained by successive outlets; the first, the "River

Warren" (discovered in 1868 by Warren[466]), was directed across the continental divide at Fort Shelling by the Minnesota valley, its original altitude being 1055 ft (*c.* 320 m) on the Minnesota–South Dakota boundary—it excavated a rocky valley, up to 4 miles (*c.* 6·5 km) wide and 300 ft (*c.* 90 m) deep, and left traceable effects for hundreds of miles farther downstream; the second was into the Laurentian glacier-lakes and so into the Mississippi, Mohawk and Hudson rivers, and later by the Mattawa River to the St. Lawrence; and lastly, a succession of outlets in north-west Ontario and north and north-east of the Patricia highland into Hudson Bay, their courses not yet fully explored but operating possibly 10,000 years ago.[467] The lake grew as the Keewatin ice melted northwards and the Patrician and Labradorean ice withdrew eastwards; it finally burst catastrophically along the junction of the two ice-sheets[468] (pl. XVIIIB, facing p. 465).

Upham traced 16 of the strandlines, including the highest or Herman beach at 1065 ft (*c.* 325 m), which were accumulated before the River Warren outlet was abandoned, and 11 of later date, together with banded seasonal clays, up to 15 m thick,[469] and deltas of enormous extent, up to 80 miles (*c.* 130 km) from back to front, which the ice built out from its edge during its retreat from the Assiniboine and other valleys.[470]

W. A. Johnston,[471] who states that there are 50 beaches, has confirmed J. B. Tyrrell's suggestion that Lake Agassiz had a second phase, "Later Lake Agassiz" or Lake Agassiz II, when the Patrician and Keewatin ice, following a partial withdrawal, blocked the northerly drainage: the advance is proved by the continuity of the highest strandlines which extend for 250 miles (*c.* 400 km), and by a conspicuous unconformity in the lake-sediments. Johnston also states that Lake Agassiz began *c.* 23,000 years ago; that its highest levels were of the same age as those of Lake Algonquin (Ignace in Ontario and Fort William on Lake Superior are nearly on the same isobase); and that at the time of the southern outlets, Lake Agassiz was 155–138 m higher than Lake Algonquin and at least 60 m higher than Lake Duluth. Soil mechanics data have recently indicated a very short period of surface drying and a very temporary disappearance of Lake Agassiz not previously reported.[472]

Keewatin ice, when most extensive, held up broad, shallow lakes against the foothills of the Rockies as is indicated by successions of terraces, e.g. above Calgary. A still larger lake lay farther north in the Mackenzie valley[473] and others in the Peace River at *c.* 2000 ft[474] (*c.* 600 m) and in the Athabasca–Great Slave Lake region[475] at *c.* 1600 ft (*c.* 488 m); the Bear, Great Slave and Athabasca lakes are lined with ancient shores up to hundreds of feet above the present levels.

Lake Ojibway. "Lake Ojibway",[476] the last of the large lakes and the most variable and short-lived of those in Ontario, was like Lake Agassiz blocked in areas now tributary to Hudson Bay (fig. 91). Named by A. P. Coleman[477] after the local Indian tribe, it originated north of the Hudson Bay–St. Lawrence divide directly north of Lake Huron, and emptied itself for a time into the Nipissing lakes and later into the Champlain Sea. It began north of the watershed as small lakes or bays at *c.* 360 m detached from Lake Algonquin with the continued retreat of the ice and uplift of the land. Falling by several stages as its marginal terraces record,[478] it became merged with a Lake Barlow[479] which filled the Timiskaming valley and had its outlet along the gorge of the Ottawa River. Widening steadily as the ice

withdrew northwards and the level fell, this Lake Ojibway–Barlow inundated an area larger than Lake Superior and covering at least 50,000 sq. miles (*c.* 130,000 sq. km)[480] between the 76th and 88th meridians, the varves forming the "clay belt" of north Ontario and north Quebec (and locally con-

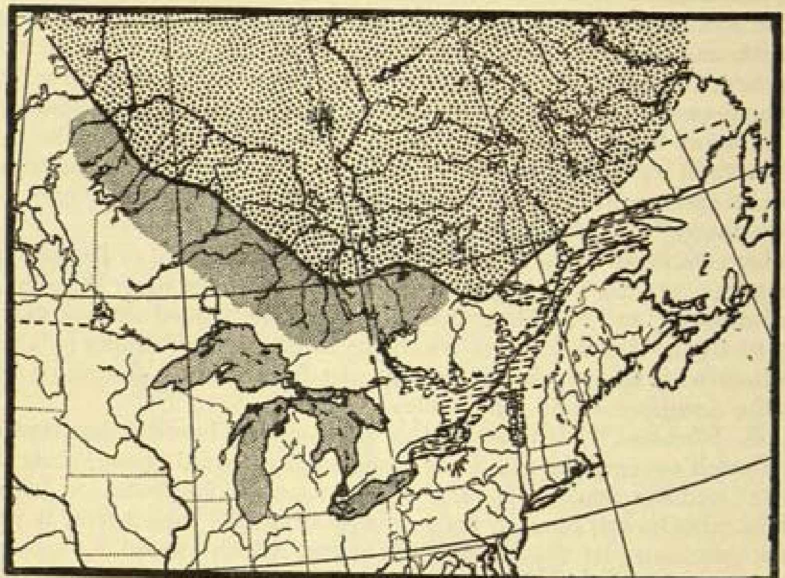


FIG. 91.—Map of Glacial Lake Ojibway, Gilbert Gulf (Champlain Sea) and Nipissing Great Lakes. Ice, large dots; freshwater, small dots; Gilbert Gulf, short dashes. A. La Roque, *G. S. A. B.* 60, 1949, p. 375, fig. 2.

taining fossil fish) and extending to the north-east 100 miles (*c.* 160 km) east of Bell River.[481] Its area and level fluctuated as the oscillating edge receded —Antevs[482] recognises three advances, one of at least 70 miles (*c.* 115 km)— but apparently finally disappeared rather suddenly when the ice stood in Quebec a little north of the line of the transcontinental railway. The waters now began to flow to James Bay across the rapidly thinning ice[483] or escaped around the ice-front to the sea near the mouths of Hayes or Nelson rivers.[484] The middle stage of Lake Barlow may have been synchronous with the final drainage and disappearance of Lake Agassiz[485]: Lake Ojibway-Barlow may for a short time have been merged with Lake Agassiz[486]: a link between the two lakes is suggested by certain fish distributions.[487]

Very little is known about the dissolution of the ice north of the Great Lakes. There were, however, several glacier-lakes (of Mankato age) in the region of James River[488] in South Dakota, a Lake Souris and Lake Dakota in North Dakota[489] (Lake Souris drained first into the Missouri River, later into James River and Lake Agassiz) and in the Hudson Bay drainage in Saskatchewan and Alberta and North-west Territories,[490] including Lake Regina which discharged into Lake Souris and later, like this lake, became confluent with Lake Agassiz. Beaches, but little studied, are common around the Athabasca and Bear Lakes and north-west of Hudson Bay[491] where they occur up to 800 ft (*c.* 244 m).

Cordilleran Lakes. Terraces with uniform level, up to 1520 m A.S.L. and associated with delicately stratified White Silts, have long been known

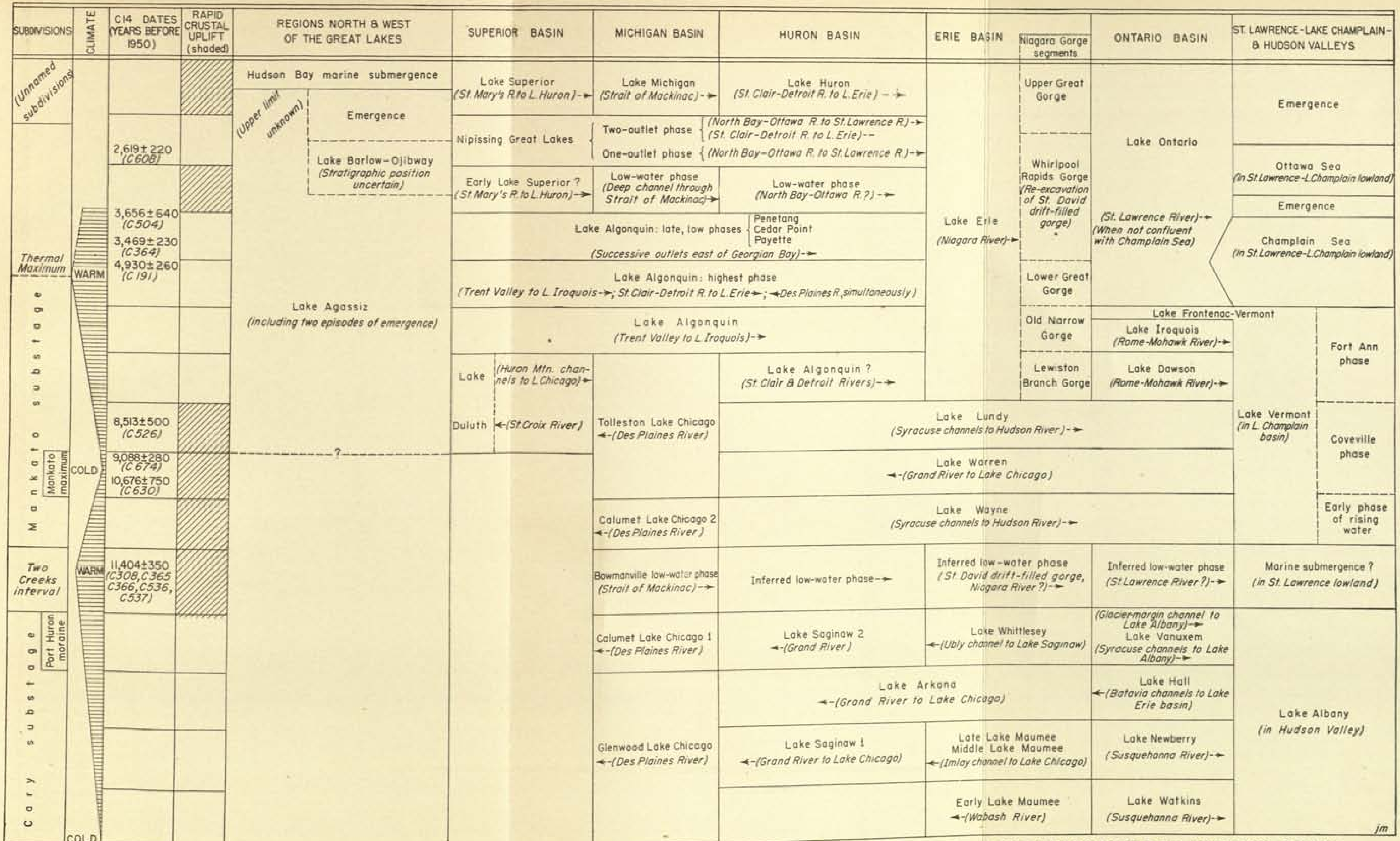


CHART SHOWING SEQUENCE, MUTUAL RELATIONS, AND OUTLETS OF THE GREAT LAKES, AND FEATURES RELATED TO THEM

To face p. 478

to be widely distributed in the Cordilleran valleys. The silts are exposed in banks up to 400 ft (*c.* 120 m) high, have a fine, uniform texture and are horizontally stratified. G. M. Dawson,[492] their discoverer, regarded them as marine. But the increasing coarseness along the margins where they pass into sands, their occasional *Unio* shells, and their coincidence with cols show that they were made by glacier-lakes[493] as Upham[494] anticipated, though R. F. Flint[495] related them to stagnant glaciers occupying the valley bottoms. The White Silts are fresh felspathic rock-flour derived by glacial grinding from the felspar of granite intrusions.

Glacier-lakes, associated with local glaciers, existed in the region of the Yellowstone Canyon.[496] The Scabland had several glacier-lakes[497] including Lake Leverett which covered about 5000 sq. miles (*c.* 13,000 sq. km) and was about 1250 ft (*c.* 380 m) deep within the Columbia Canyon.

2. Glacio-marine Deposits

Glacio-marine sediments are laid down very rapidly if ice or its streams transport material into the sea[498]—the delta in Seal Bay in front of the Hidden Glacier of Alaska advanced *c.* 490 m between 1899 and 1910 at a maximum rate of *c.* 69 m/annum.[499] They differ from glacio-lacustrine deposits because wave-work is more powerful, flocculation prevents varves from forming, and tides cause a constantly fluctuating sea-level. Although the muds are deposited within a few kilometres of the entrance of the streams,[500] as in Spitsbergen or Alaska, and give rise to the *örer* (plural) which appear in many Norwegian place-names (see p. 505), they may be so vast that they fill the whole fjord, as in Norway where they form the *Fjordmjäle*, and in Iceland[501]—the strand north of Katla has in historic time been pushed back 2 km and the Kerlingarfjord of the Vikings is now entirely replaced by a sand plain. They build flat, sandy or clayey expanses at the heads of ice-fjords in Greenland,[502] and their higher parts are passable only in shallow boats. A stretch of 50 km may be partially laid dry; these are the *isortoq* or "clay fjords".[503] All stages of infilling are recognisable; they range from fjords in which the clay covers the bottom to others where rivers meander either in innumerable braided courses across a naked flat, the Greenlanders' *narssak*,[504] or, following an uplift, between a flight of terraces of variable width or in canyons incised in the plain.[505]

Ross Sea has a carpet of stiff yellow clay, rock-flour milled by glaciers on South Victoria Land.[506]

Extensive deltas have been described from the mouths of modern glacier-streams in the Arctic[507] and from Pleistocene shore-lines in Fennoscandia[508] and Maine.[509] The huge amount of detritus furnished by the melt-waters of the Glacial period has helped to build the Mississippi delta, though fluvio-glacial deposits of coarse texture are scanty in the lower reaches of the river.[510] In general however such deltas were confined to protected bays or to openings between masses of stagnant ice.[511]

Glacio-marine sediment contains little life except at the shore.[512] In deeper waters they form pell-mell gravels and clays, with no further sorting or abrasion, while at shallow depths the gravel, sand and silt grade from coarse to fine as the off-shore depth increases. In Baffin Bay, the deposits are apparently coarser than in the Greenland fjords.[513] In exposed localities, as before the Malaspina Glacier, where waves, tides and currents are active and supreme, the deposits are marine rather than glacial.

It is difficult in an uplifted accumulation of glacio-marine sediments to fix precisely its former intersection with the water-surface. It probably coincided with the lower limit of cobbles and gravel, the subaqueous layers being finer and more uniform and containing some mud. The difficulty is enhanced if residual ice in an estuary permitted the plain to grow above sea-level.[514]

Ice standing with its front in the sea does not build definite moraines, since currents or bergs transport its material and its floating continuation responds readily to oscillations.[515] The frontal deposit is smooth and level and much weaker topographically than the corresponding land accumulation. Its structure is generally deltaic, the beds grading into thinner and finer sediments and interlaminating with marine clays containing possibly marine shells. Such frontal accumulations, often noticed about modern polar glaciers,[516] comprise most of the Finnish and Norwegian marginal deposits of lateglacial date, laid down in the then sea.

The deltas have steep ice-contact proximal faces and distal frontal slopes (fig. 92). Their texture and structure differ from the marginal deposits that

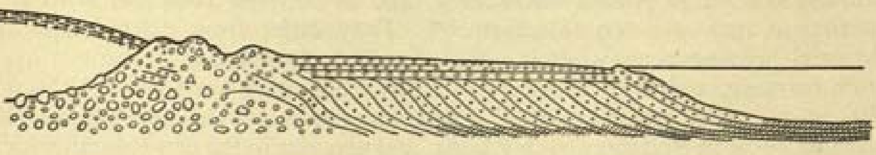


FIG. 92.—Diagrammatic section through a fluvioglacial marine delta. M. Sauramo, *1478*, p. 9, fig. 1.

were originally laid down on land and later submerged, e.g. the Horn's Reef of the North Sea which is the retreat moraine of a penultimate glaciation of Denmark,[517] or those sometimes discovered on the bottom of the southern Baltic[518] or off the Alaskan coast[519] or that rise as islands or submerged banks and shoals in the fjords of Greenland,[520] in James Bay,[521] off the coast of Maine[522] (these are probably submerged cuestas[523]) or possibly on the continental shelf of west Scotland[524] or west Norway (see p. 1172).

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1888, 199. **435** *A. J. S.* 141, 1891, 12; 143, 1892, 210. **436** *372*, 43; G. M. Stanley, *G. S. A. B.* 47, 1936, 1933. **437** *623*, 42. **438** *623*, 36; G. K. Gilbert, *Am. G.* 18, 1895, 231; J. W. Spencer, *op. cit.* 198. **439** *298*, 31, 51; *863*, 14. **440** *372*, 87. **441** G. M. Stanley, *G. S. A. B.* 47, 1936, 1951; 48, 1937, 1680. **442** H. L. Fairchild, *N. Y. Mus. Rp.* 54 (1), 1900, r. 106 (lit.), pl. 19. **443** *863*; J. W. Goldthwaite, *G. S. A. B.* 22, 1910, 227. **444** *298*, 71. **445** *298*, 70; A. P. Coleman, *CR. C. G.* 1914, 441. **446** *298*, 48. **447** *612*, 67; A. P. Coleman, *G. S. A. B.* 15, 1904, 347; H. L. Fairchild, *N.Y. Mus. B.* 127, 1909, pl. 42; 160, 1912, pl. 17. **448** *491*, 215. **449** M. B. Baker & A. W. Johnston, *T. R. S. Can. 3*, 28, 1934, 75; F. B. Taylor, *Sm. I. Rp.* 1912, 291. **450** *Ibid.*, *Am. G.* 17, 1897, 65. **451** *1002*, pl. 8; *1005*, 447. **452** *1002*, 69. **453** G. M. 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B.* 906, 1939, 65. **490** J. Bird, *A. J. S.* 251, 1953, 215; W. A. Johnston & R. T. D. Wickenden, *T. R. S. Can. 3*, 24, 1930, 41; 25, 1931, 29, fig. 1; M. Y. Williams, *ibid.* 23, 1929, 70. **491** D. A. Nichols, *ibid.* 26, 1932, 270; J. B. Bird, *G. S. A. B.* 65, 1954, 459, 462 (map). **492** *369*, 36; *Can. Rp.* 7, 1896, 276B, 290B. **493** R. A. Daly, *Can. Mem.* 68, 1915, 149; S. Duffell & K. C. McTaggart, *Can. Mem.* 262, 1952, 71; W. H. Matthews, *T. R. S. Can. 3*, 38, 1944, 39; J. B. Tyrrell, *G. S. A. B.* 10, 1899, 197. **494** *1735*, 208. **495** *T. R. S. Can. 3*, 29, 1935, 107. **496** A. D. Howard, *G. S. A. Spec. P.* 6, 1937, 81 (lit.). **497** *774*, 21, 22 (map). **498** W. H. Brewster, *A. J. S.* 129, 1885, 4. **499** *1681*, 159. **500** *1681*, 465; G. De Geer, *Y.* 33, 1931, 148. **501** *1704*, 76; T. F. W. Barth, *Carn. I. P.* 587, 1950, 11; R. Brown, *Q. J.* 26, 1870, 682; T. Thoroddsen, *D. Vid. Selsk. 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LA.* f. 1904, 377. **519** *1084*, 328; *1681*, 358; U. S. Grant & D. F. Higgins, *U. S. G. S. B.* 526, 1913, 44; L. Martin, *P. M.* 1912, II, 149. **520** *114*. **521** A. P. Low, *G. S. A. B.* 4, 1893, 420. **522** W. Upham, *P. A. A.* 20, 1879, 221. **523** *861* **524** O. Holtedahl, *T. E. G. S.* 15, 1952, 220.

Chapter XXIV

GLACIAL EFFECT ON SCENERY

General. The Pleistocene period has had an economic value in providing supplies of water and of building and other materials, including rich auriferous and other metalliferous placers. It has initiated lakes and waterfalls for power, transformed the soils (see p. 359), controlled agriculture and influenced modern culture, communications and settlements. The ice has left its stamp upon the relief, has enriched the world by the addition of alpine scenery, and has complicated the geomorphological problems no less than those of the distribution of organisms.

The major features in glaciated regions reflect the preglacial configuration in all essentials. The broad distinction between hill and valley had already been asserted; mountains and lands had their present values in altitude, mass and general slope. Ice in other words did not originate so much as modify. Where it overflowed a country, it rounded and softened the outlines, ground off the inequalities and subdued any ruggedness. It sensibly lowered the hills, freshened and steepened the scarps, and carried away their talus; it broadened and deepened the valleys that ran parallel with its flow and spread the detritus over the lower ground. The general shedding of waste from the colder higher surfaces to the warmer levels of the valley floors has been of great human benefit, e.g. in the Andes.[1]

In the aggregate, ice (and water working under conditions imposed by it) has left an important legacy and exerted a not inconsiderable scenic and cultural influence. In erosional centres its modifications were frequently profound (see chs. XII–XV). It heightened the topographical contrasts, sharpened peaks, narrowed divides to arêtes, lowered cols to comfortable passes, deepened valleys, converting them in places of great ice-concentration into through valleys, and imparted a wild and savage scenery. It multiplied the number of waterfalls, creating those where the waters plunge over the walls of cirques or U-valleys, saw their way across rock-barriers, or run through glacial epigenetic gorges or rapids, the *koski* of Finland. The great development of water-power is an important legacy of the Glacial period.

Ice often intensified the influence of rock-structure, especially if, as in the Laurentian Highlands,[2] this was diverse or if igneous and sedimentary rocks were in contact.

In zones of deposition, glaciation has almost entirely controlled the detailed relief and drainage. It has evened up the surface, buried or removed minor features, and filled in old valleys in varying degrees according to their relation to the ice-flow. Even where, as in supramarine Finland, only 2% of the area is occupied by fluvioglacial material, this has exercised a disproportionate influence[3]; the clays and sands in their separate distributions have even influenced the distribution of the different types of plough. Over extensive tracts, as in the plains of Cheshire, East Anglia and north Germany and in the Mississippi valley, bedrock, deeply hidden beneath drift, is unable to assert itself. New surfaces have been built up, with new drainage systems

and new divides which sometimes pay scant regard to the ancient patterns. These contrast with the central areas in which the outcrop of bare rock has been much enlarged and glacial accumulations seldom conceal earlier features. Much of Canada has been scoured clean of its weathered debris so that fresh rock has been exposed for effective prospecting and mining.

The drifts have also influenced the vegetation. For example, in the British Isles[4] the sands and gravels bear heath or corresponding woodland, the non-calcareous boulder-clays oakwood, and the chalky boulder-clay ash-oakwood.

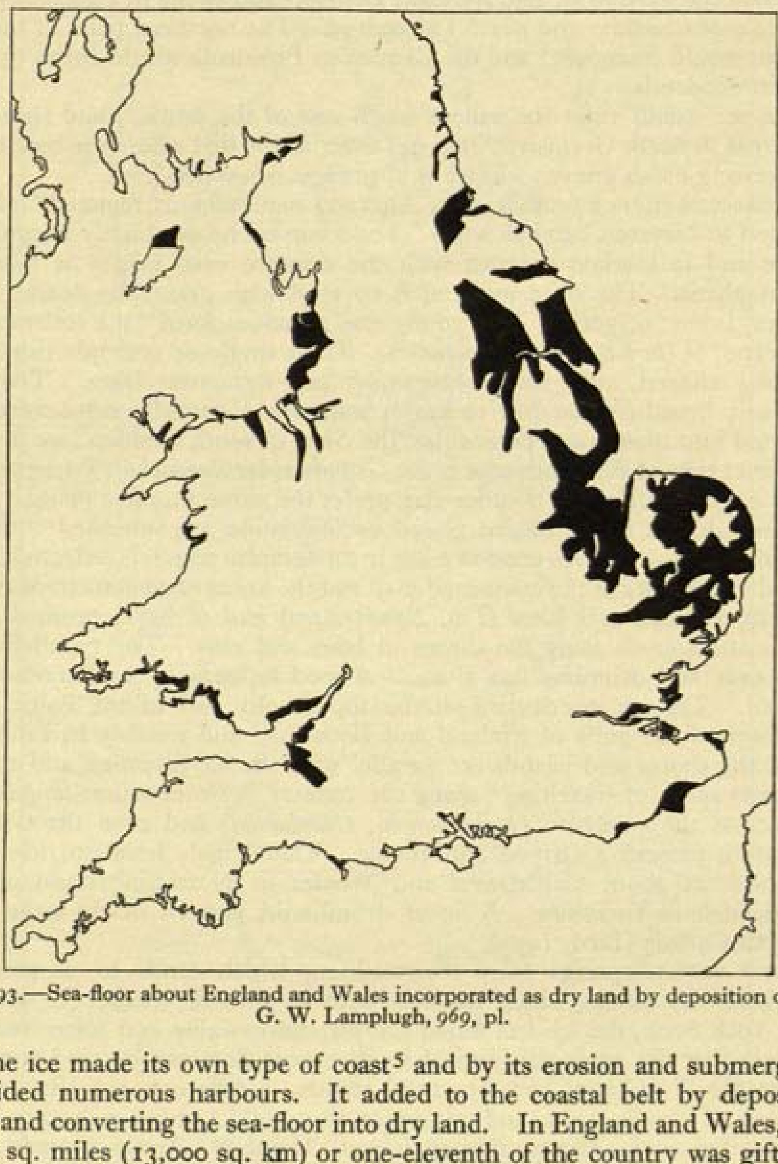


FIG. 93.—Sea-floor about England and Wales incorporated as dry land by deposition of drift
G. W. Lamplugh, *969*, pl.

The ice made its own type of coast[5] and by its erosion and submergence provided numerous harbours. It added to the coastal belt by depositing drift and converting the sea-floor into dry land. In England and Wales, over 5000 sq. miles (13,000 sq. km) or one-eleventh of the country was gifted in this way[6] (fig. 93). Were the drifts removed, the sea would spread inland from the present coasts of Northumberland and Durham, especially in the valleys of the Tyne and Tees. It would flow up the Vale of Pickering as far

as Malton and attain its old shore-line where chalk cliffs swept west of Holderness as a bold curve from Bridlington to Hessle and continued south of the Humber. The Vale of York would be submerged as far as Boroughbridge, the Vale of Trent to the Lincolnshire Wolds, and the River Witham to Lincoln: north-east Lincolnshire would be an island. The Fens would be hollowed out into a big bay and roughly half of Norfolk and Suffolk would be inundated.[7] The Vale of Clwyd as far as Ruthin and the Dee nearly to the entrance to the Vale of Llangollen would constitute arms of a sea which would also flood the western strip of the Lake District[8] and sweep in a great bay over the plain of Cheshire and north Lancashire. The northern plain of the Isle of Man would disappear[9] and the Carnarvon Peninsula would break up into a chain of islands.

The sea would enter the valleys south-east of the Baltic[10] and submerge vast areas in north Germany[11] (fig. 94) as far as the line where the base of the drift resting on an uneven surface is at present below sea-level.

Pleistocene stony moraines in the Alps and mountainous regions, "hillocks dropped in Nature's careless haste", build conical heaps, highly irregular in profile and in marked contrast with the smooth, even slopes or boulder-strewn plains. The finer material is covered with grass, the coarser with heather, broom or gorse. The young end-moraines form "the end-moraine landscape"[12] (*la passage morainique*[13]). Their single or multiple ridges are variously shaped, with wavy topography and numerous lakes. They are extremely broad (in the Inn 10 km[14]) and, where partially submerged, are dissected into islands and peninsulas, the *Skier* of south Sweden (see p. 494). A distinct type of this landscape is the *Grundmoränenlandschaft*[15] (supporters of the englacial origin of boulder-clay prefer the name moraine plain).

Scenic details in regions of glacial accumulation are inherited from the Glacial period. The ice created a linear topography which is reflected in the striped appearance of the contoured map and the linear arrangement of ridges, marshes, streams and lakes (Fin. *Seenstrassen*) and of bays, promontories, islands and sounds along the shores of lakes and seas. The parallelism of lakes, osar and drumlins has given a striped topography to Fennoscandia (fig. 96). The ice inscribed a similar topography east of the Baltic,[16] on the floors of the gulfs of Finland and Bothnia,[17] and notably in Finland[18] where the shores and islands are parallel with the ice-direction and the inhabitants speak of travelling "along the country" (*pitkinmaisin*; *långslandet*) or "across the country" (*poikkimaisin*; *tvärslandet*) and even the types of vegetation present a striped appearance. The British Isles provide other examples, as about Coldstream and Wooler in Northumberland and in Wensleydale in Yorkshire. A linear drumlinoid pattern marks large areas of north Canada (Bird, 1953).

Some fluting springs from ice-moulding which tends to deepen and straighten the valleys which conform with its own direction. In parts of New York State, the ice has fluted the ground-moraine and softer rocks so strongly that the principal lines of its divergent flow can be read from the topography.[19] Glacial and structural trends, if different, are often superimposed; if equally strong and transverse, they build a peculiar checker-board topography, as in Mount Desert Island, Maine.[20] Arctic Canada often displays a similar linear topography.[21]

Lakes. Glaciation has profoundly altered the drainage. The uncertainty and immaturity it induced, well seen for instance in Finland where the

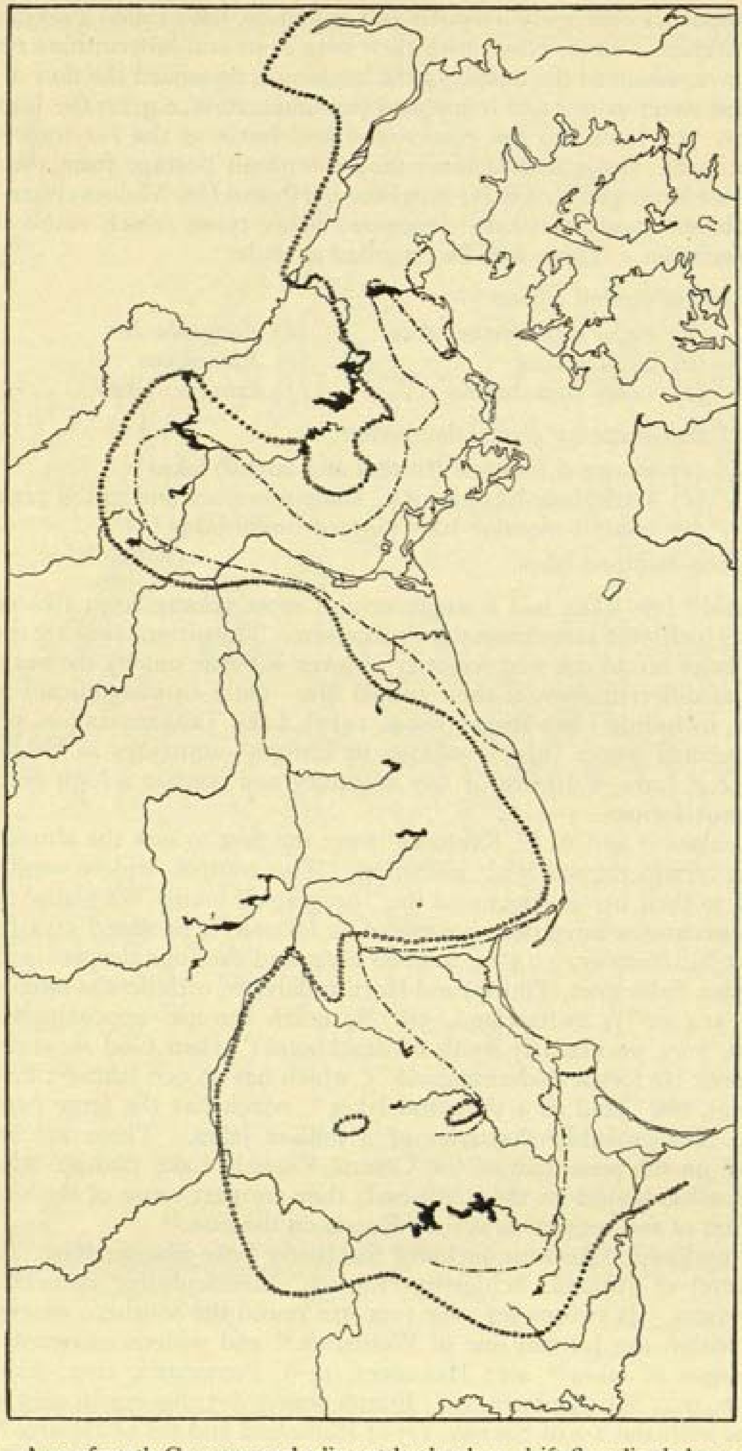


FIG. 94.—Area of north Germany and adjacent lands whose drift-floor lies below sea-level. Lower surface of the drift at sea-level (circles) and at −50 m (dot and dash). K. Buerlen *et al.*, *Z. Gl.* 21, 1933, p. 45, fig. 5.

water-sheds are extremely irregular and indefinite, have called a myriad lakes into existence. These lakes, with their long arms and labyrinthine ramifications, have enhanced the beauty of the landscape, equalised the flow of rivers, furnished water-power and influenced communication, e.g. for the lumbering industry. In Iowa too the canoes and keel-boats of the fur-traders from Canada to St. Louis and Orleans moved without portage from the Upper Minnesota River and Red River into Blue Earth and Des Moines rivers.[22]

All classifications of lakes[23] recognise many types which result directly from glaciation. These may be classified as under:

1. Lakes eroded by ice
   - (a) roche moutonnée lakes
   - (b) cirque-lakes
   - (c) valley rock-basins
   - (d) *Zungenbecken*
   - (e) *Rinnenseen*
   - (f) eversion lakes
2. Lakes made by glacial deposition
   - (a) marginal moraine (frontal and lateral) lakes
   - (b) kettle-hole lakes in osar, kame-moraines and pitted plains
   - (c) ground-moraine lakes; interdrumlin lakes
3. Ice-dammed lakes.

Probably few lakes had a single origin: most sprang from two or more actions which were simultaneous or successive. Transition types are common.

The brief life of the vast majority of lakes is borne out by the very feeble degree of differentiation of their animal life. Only an insignificant number of lakes, including Lake Baikal (see p. 1417), Lake Tanganyika (see p. 1417), Lake Ochrida (see p. 1417)—unique in Europe—and lakes in Celebes and China (e.g. Lake Tali) are of any antiquity and contain a high percentage of endemic forms.

F. Leblanc[24] and A. C. Ramsay[25] were the first to link the abundance of lakes in certain regions with glaciation. The control, widely confirmed,[26] has led to their being designated the "orographic fossils" of glacial geology. Lake-percentages have been computed as follows[27]: Scotland 1·12 (Sutherland 3·5[28]); Norway, 3·14 (*c.* 200,000 lakes and the four deepest in Europe, viz. Mjösa, Salsvatnet, Tinnsjö and Horninsdalvatn, with depths ranging from 443 to 514 m[29]); Switzerland, 3·67[30]; north Europe, approximately 4[31]; Sweden, 8·03, particularly south of Stockholm ("when God separated land from water He forgot Södermanland"), which has 96,000 lakes[32]; Finland[33] or Suomi, the "land of a thousand lakes", which has the large percentage of 19·9 and probably a quarter of a million lakes. These are specially plentiful on the peneplain of the Central Finnish Lake plateau which the Salpausselkäs bound on the south-east: they are part cause of the sharp demarcation of the vegetation against Russia on the east.[34]

The peribaltic lake-zone includes the Baltic Lake-plateau (Ger. *Baltische Seenplatte*) of Jutland, Schleswig-Holstein, Mecklenburg, Pomerania and East Prussia. It sweeps for over 1200 km round the southern shores of the Baltic within the Jutnian line of Woldstedt[35] and widens eastwards. The percentages of lakes[36] are: Hannover, 14·6; Pomerania, 10·2; Schleswig-Holstein, 9·3; Mecklenburg, 9·2; Brandenburg, 8·7; figures in very striking contrast with the 2·2 of Saxony, 1·7 of Rhineland and 0·1 of Hessen-Nassau.

A map[37] showing the relation of the bigger lakes of North America to glaciation brings out the same control (fig. 95). For example, the

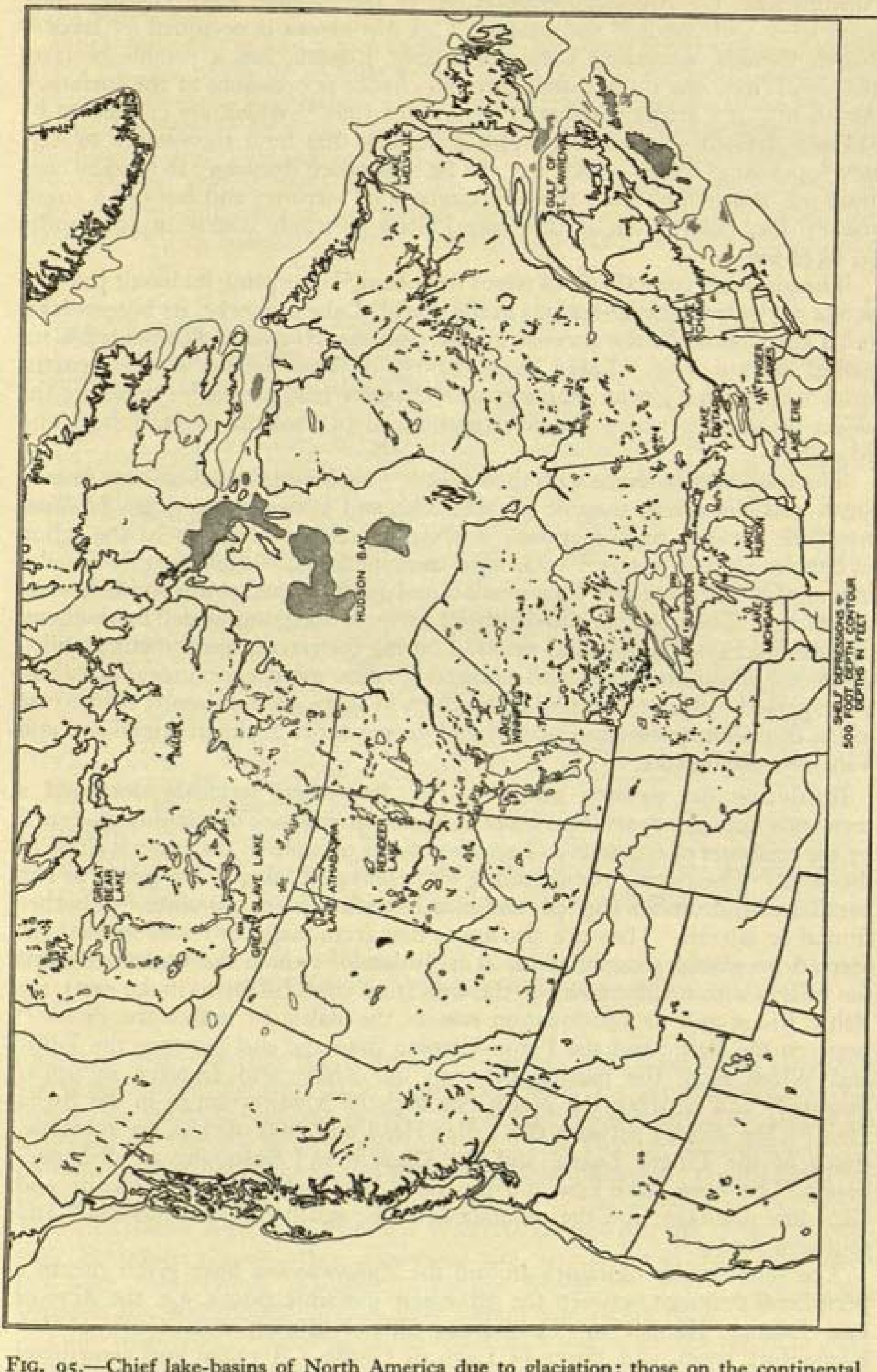


FIG. 95.—Chief lake-basins of North America due to glaciation: those on the continental shelf are shaded. F. P. Shepard, *J. G.* 45, 1937, p. 79, fig. 1.

Adirondacks, the American counterpart of the English Lake District, have 2000 lakes and ponds,[38] and nearly 7% of Minnesota is occupied by lakes.[39] North Canada, including Labrador's Lake Plateau, has a jumble of lakes that spill from one to the other through chance depressions of the surface.[40] About 6% of Canada consists of freshwater-lakes[41] which are connected by sharply defined river-channels and waterfalls that lend themselves to easy development of water-power, e.g. the St. Lawrence drainage: in marked contrast the Mississippi flows through unglaciated territory and has not a single natural lake (apart from ox-bow lakes) along its whole course of 4000 miles (*c.* 6450 km).

The glaciated coastal belt of west Greenland,[42] excepting its basalt portion, is one of the richest lake-regions in the world; Lake Giesecke, its biggest lake, is 50 km long while the Steenstrup lakes on the Nûgssuaq Peninsula are together 70 km long. Lakes cover 10·6% of Bear Island.[43] The partial drowning of such glacier plains gives a characteristic littoral[44]: these are the *skjärs* of Sweden which are also reproduced in Finland, west Ireland and Maine.

The lakes vary in shape with their origin. End-moraine lakes[45] are usually small and elongated; they lie parallel with and between the ridges (*Faltenseen*[46] lie between push moraines) and were liable to disappear by the action of lateglacial melt-waters.[47] Ground-moraine lakes,[48] often roughly circular in plan (Ger. *Beckenseen*[49]), are shallow and gently sloped above and below the waterline. Their outline and depth are irregular and complicated by shallows, islands and bays: fluvioglacial material during the retreat has sometimes filled them in. Drumlins have also clogged a once mature drainage; they are interspersed with interdrumlin lakes which in shape are frequently the reverse of the drumlins, possessing a rocky and deeper shore at the proximal end and a shallow distal shore.[50]

Influence on stream courses. Ice has often seriously deranged a country's drainage apart from lakes. Minor streams are commonly governed by the contours of the drift and are sometimes oblique to the general slope of the land. The rivers, often swollen into elongated lakes,[51] are prone to run parallel with drumlins (fig. 96) and osar or with lines of moraines,[52] whether frontal or lateral. They are also apt to flow from major or minor watersheds erected by glacial accumulation, as in Jutland[53] (where the watershed parts the valleys with mature forms in the west from youthful forms in the east), the Baltic Ridge and its continuation east of the Baltic (it makes the divide[54] between the Baltic and the Ponto-Caspian drainage and between the Baltic and White Sea), the parting between the Rhine and Danube in upper Swabia,[55] and between the North Sea and the Mediterranean in the Swiss Plain[56]; the divides between the Valley Head moraines of the Cary substage, south of the Finger Lakes, and the Ontario and Susquehanna drainage-basins,[57] between Lake Erie and the Ohio, and between the Missouri and Canadian drainage; and the watersheds in the north Siberian tundras[58] and southern Andes.[59]

The semicircular moraines around the *Zungenbecken* have given rise to a peripheral drainage between the successive morainic ridges, e.g. the Arve of Lac Léman, Kander of Thunersee, Sill of Zürich See, Gurk of the Klagenfurt basin, and Sarca of Lago di Garda. A radial and centripetal flow occurs towards the inner basin, e.g. around Boden See,[60] making use of hollows between the moraines and drumlins and the axes of the *Zweigbecken*.

Centripetal drainage characterises the peripheral diffluence terrain on the ice-ward side of cols.[61]

While the coming and going of the ice-sheets have made little difference where the flow of the ice and the natural drainage were in the same direction, streams elsewhere have often been deflected into channels excavated by marginal waters[62]: the Missouri shows many instances of such diversions by the Kansan and Illinoian ice (see below). Their anomalous courses are expressed by abrupt turns and striking detours, by abandoned valleys of normal form, and by a disproportion between valleys and their occupying streams. Rivers have often pieced together old and previously independent

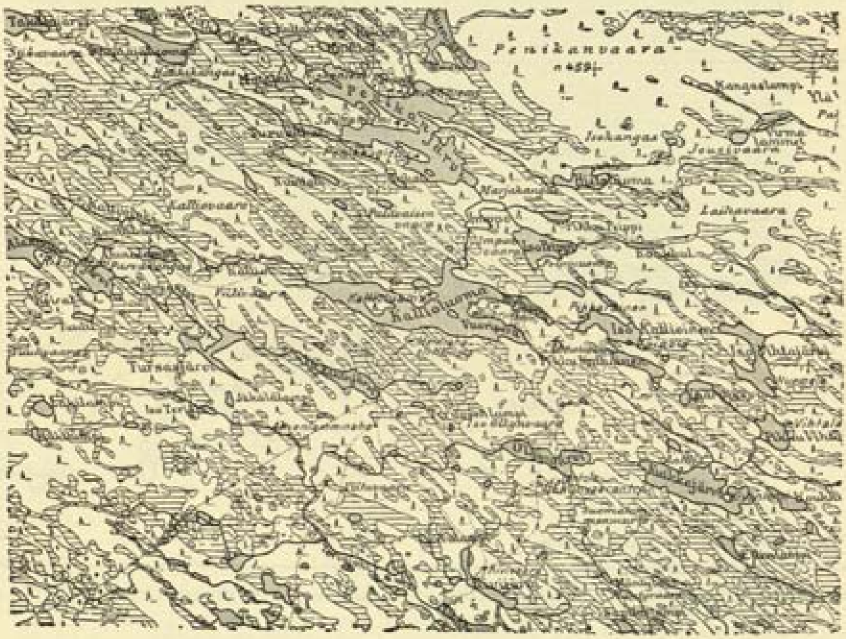


FIG. 96.—Striped appearance of drumlin field of Kuusamo, Finland; ruled areas, peat bogs; grey, lakes. K. Virkkala, *1920*, p. 57, fig. 26.

courses. Big rivers like the Vistula, Oder, Spree and Havel follow the north German *Urstromtäler* over parts of their course and are permanently diverted westwards along the zig-zag paths. Fluvioglacial aggradation by alpine ice deflected the Danube in its upper waters.[63] The Thames was expelled from its more northerly course in two stages of glacial diversion; the first was the work of a local Chiltern ice-cap, the second of the advance of the Great Eastern Glacier deploying from the Wash region (see pp. 768, 997). North-east Yorkshire shows other diversions (fig. 97) and the main watershed has been considerably modified in central England[64] where virtually the whole of the Warwickshire Avon river system is of recent date.

The Pleistocene vicissitudes of the Rhine and Meuse are not yet all known. Nevertheless, moraines, outwash and subsidence turned the Rhine from its earlier way to the Zuider Zee[65] along the Ijssel and Gueldsche Poorte from

above Kleve, and shifted the Meuse westwards about Maastricht.[66] The two rivers suffered various changes during the Glacial period.[67]

Three great North American rivers owe part of their situation to diversions around the ice. The Allegheny–Ohio represents the union of formerly distinct valleys[68]: this great access of headwaters has favoured the cutting power of the river in its lower reaches. The upper Missouri and its tributaries as far down as the White River in South Dakota appear preglacially to have drained northwards into the Red River basin and Hudson Bay instead of to the Gulf of Mexico as to-day (see p. 286). The Keewatin ice, of Illinoian age, diverted the drainage through the Dakotas to the west over a stretch of 155 miles (*c.* 250 km), the old valley being mostly deserted.[69] The valley of the upper Mississippi was similarly repeatedly overspread by ice.

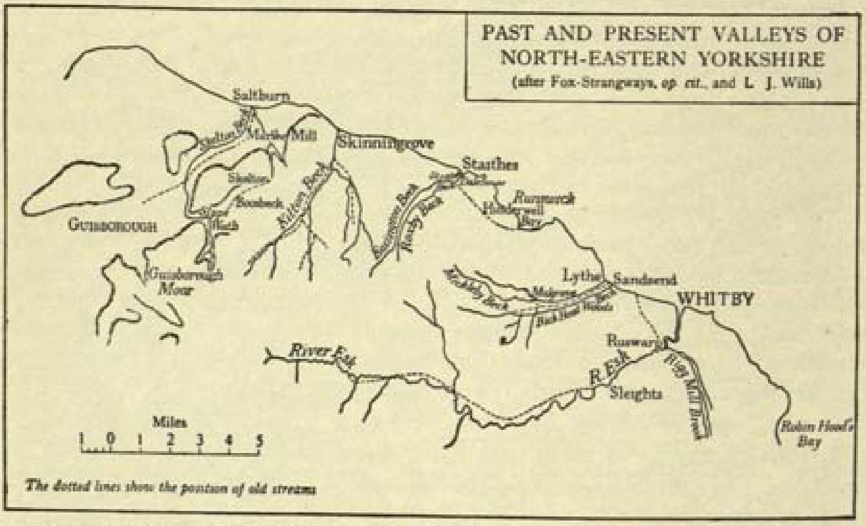


FIG. 97.—Past and present valleys of north-east Yorkshire. J. A. Steers, *1624*, p. 466, fig. 101.

During each invasion, the valley was mutilated or obliterated and the river was pushed from its course to seek new channels.[70]

The suggested postglacial age of the Yellowstone Canyon is apparently without foundation, for there have been no radical changes of a permanent character since the Yellowstone drainage was initiated.[71] Equally untenable is the suggestion that the Colorado River began about the beginning of the Pleistocene.[72] It is, however, possible that by erosion and deposition under Pluvial conditions the interior basins of the Niger, Nile, Zambesi and Hwang-ho were converted into oceanic drainage.[73]

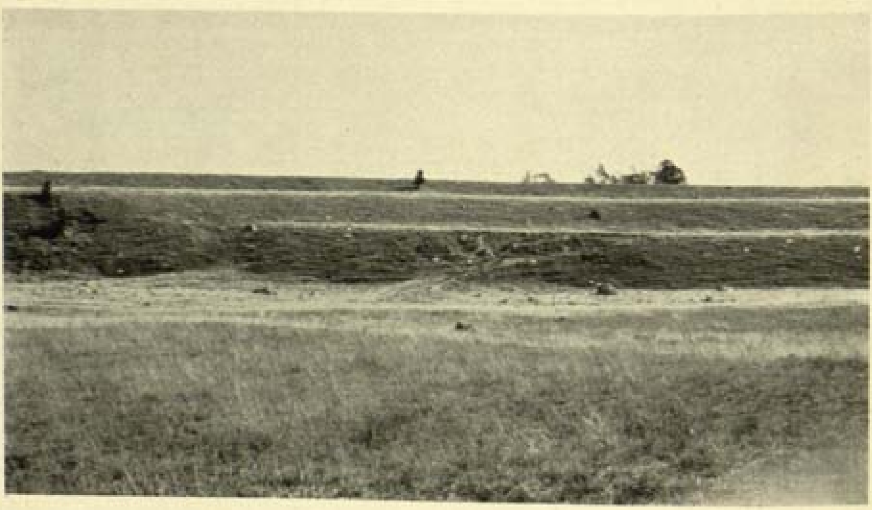
There have been diversions by outwash and moraines,[74] particularly of tributaries along the outer edge of lateral moraines, and by terraces, notably those which lie at a critical level below the intake of spillways across cols ("pseudo-cols"[75]), e.g. at the head of the Finger Lakes streams, where moraines have turned the waters southwards.[76]

The complex changes each glaciation wrought are to be seen in the disarranged drainage. Glacial epigenesis, rare in the thin and unobtrusive drifts near the glacial limits and of little account in mountainous areas (unless thick

A. Delta of Farmington River built into glacial Lake Farmington, Connecticut [R. J. Lougee]

B. Overwash delta (*on left*) of the Irish Sea ice in glacier Lake Ennerdale, Lamplugh, Cumberland, with contact slope and kettle-hole in foreground
[Geol. Surv. Gt. Britain : Crown copyright]

A. Parallel Roads, Glen Roy, Inverness-shire, Scotland
[Geol. Surv. Gt. Britain : Crown copyright]



B. Strandline of Lake Algonquin, Waubaushene, Ontario
[Geol. Surv. Canada : Crown copyright]

(facing p. 497)

drift in the main valley has displaced the tributaries[77]), is a very common heritage on plains, as in Finland,[78] south Bavaria and the Swiss Plain,[79] or over much of Canada and the Arctic Archipelago,[80] or in those preglacial valleys and adjacent tracts which were buried under an extensive pall of drift. Rivers, developing by stages as the ice uncovered the land, automatically resumed their flow over their "false floors", following the lowest levels of the reconstructed surface. In most cases, the preglacial and postglacial drainage lines tend to coincide, especially if the ice flowed along the major valleys and the postglacial streams followed sags above them. The present occupants then combine ancient and modern erosion contours, hemmed in between banks of drift, sometimes touching the old valley at the bends as they swing across it. The case is different if the relief is slight, as on the even palaeic surface of Norway[81] (see p. 348), if the preglacial valleys were transverse to the ice-flow (see p. 323), or if the rivers flowed towards the source of the ice, i.e. in general northwards in north-west Europe and North America, or are constricted as in many rejuvenated valleys, e.g. in the glint country east of the Baltic.[82] The rivers are then apt to wander out of their ancient valleys or cross them diagonally, excavating gorges on one side or the other or touching them at the elbows or turns: the earlier pattern may be wholly lost as in northern Ohio, Indiana and Illinois where even the general plan and direction of the preglacial valleys are unknown.[83] Such diversion valleys have often been used for the sites of dams of reservoirs.[84]

Hence postglacial rivers often betray a lack of harmony between their parts. Open stretches over broad, drift-choked valleys alternate with narrow, rocky gorges, the valley shape changing rapidly or suddenly. The drift-free gorges are almost always shallower than their clogged or half-buried predecessors and little if at all wider than the occupying streams: the shape and lack of any appreciable bevelling of the edges attest their newness. Terraces in the expanses above the gorges register stages in the downcutting.

Plugged valleys are revealed by bores for economic purposes, as in coal-fields. The filling is glacial sand and gravel, silt or till: it constitutes the "sand bars", "gravel faults" and "clay wash" of English miners and the "sand and clay dykes" of Scotland. The gravels often form most prolific sources of underground water, e.g. in Illinois.[85]

Epigenetic gorges are apt to occur, e.g. in New York State,[86] on one particular side of an ancient valley. This may be because that side was sunnier[87] or winds displaced the deltas and rivers in that direction[88] or because deposition was more active in the lee of the iceward wall and virtually absent at the impact side,[89] e.g. in preglacial valleys athwart the ice-flow. Two parallel streams may occupy an old valley in which the retreating ice left the drift with a convex surface.[90] Epigenetic gorges across preglacial meander-cores were sometimes initiated by lateral glacier-streams.[91]

Although the preglacial drainage within the limits of glaciation is but imperfectly worked out, countless examples of glacial epigenesis have been discovered. They are the rule rather than the exception in the country of low relief of south Finland[92] and are found in Russia—the Volga is pieced together of stretches of several ages.[93] Diversions also occur in Switzerland[94]—the finest is furnished by the infilled valley of the Rhine along the line Lingen-Thaingen-Waldshut and the Falls of Schaffhausen[95] (24 m high) which are cut into the hard Jura Limestone and have raised the level of Boden See by 30 m; in the British Isles, including the broad, drift-filled valley of the Team

Wash[96] that falls from Durham to the Tyne at Dunston, the Dee valley,[97] and examples in the Cleveland Hills (fig. 97), in the Lake District[98] and in Ireland[99]—these embrace the diversion of the Bush from Ballycastle to Bushmills, the Liffey from the Barrow to Dublin Bay and diversions in the Corrib and Lower Erne.

North America provides numerous instances[100]—at least six manufacturing cities on the Merrimac alone owe their sites to water-power at such points.[101] The waters of the upper Susquehanna and its tributaries originally passed through the Finger Lakes region into Ontario, and the upper Allegheny, with the whole of the north-flowing drainage of the upper Ohio, has been turned southwards from the Erie by ponding and the silting up of the preglacial valley.[102] The drainage pattern in South Dakota has been changed considerably[103]: it is shown in abandoned major valleys, in abrupt variations in the width of valleys, in anomalous upland slopes and in former major divides. Each glacial epoch produced changes in Ohio.[104] The River Niagara is the classic example (see p. 1518) and Great Falls in Labrador provides the finest spectacle in North America.[105]

In the long preglacial cycles, the drainage in lowlands had gone far towards finding rocks of weak resistance where option was allowed and in grading their courses where it was necessary to cross hard rocks. The features due to glacial derangement, e.g. lakes, swamps, falls and gorges, affect the landscape to a degree disproportionate to the change in relief.

Diversions and captures resulted from the epiglacial tilting of the lands, as in Fennoscandia[106] (see p. 1317), and from glacial erosion, as by the sapping of the heads of valleys in New Zealand.[107] Transfluence and diffluence (see p. 332) have displaced the watersheds,[108] e.g. on the Majola Pass, and have led to the capture of streams after the retreat by direct glacial lowering of the pass or by lateglacial fluviatile erosion.[109] This is particularly prone to happen if moraines or outwash lie across the earlier courses and diverted the streams across lowered cols.[110]

Cultural influence. Some of the cultural influences arising from the Glacial period have already been noticed incidentally. A few others may now be briefly recorded. In some cases the ice has made communications easier. It has produced transfluence and diffluence cols (see p. 332) which have, for example, directed human intercourse in the Alps,[111] since peoples have tended to expand outwards over transfluence passes as the ice had done previously.[112] Thus the glacis on the farther side, important for communications, has influenced historical development, political and language distributions, and has given rise to contested frontiers. The long and straight *Rinnenseen* guide roads and railways in parts of north Germany, and the terraces of the Rhine rift-valley carry the main roads through the valley as they have done since Roman times.

Abandoned marginal channels in Alaska offer an easy passage up a glacial valley for both man and beast, and similar overflow valleys, with their remarkably uniform grades, facilitate communication by road, railway or canal in both Germany and North America.[113] Thus the north German *Urstromtäler*, which in the Middle Ages bore many towns on their flanks,[114] later conducted railways between the Rhine and Vistula on their broad, flat floors (their gradient is 1:8000[115]) and canals, e.g. the Friedrich-William between the Oder and Spree, the Plauesch between the Spree and the Havel and Elbe, and the Finow, Ruppiner and Rhine canals linking the Oder with the Havel.

In North America the overflow valleys have facilitated boat traffic from Ottawa to Georgian Bay, from the St. Lawrence and Lake Ontario to the Hudson, from Lake Erie to the Wabash and Ohio, and from Lake Michigan to the Mississippi.

In other cases, communication has been hindered.[116] For example, valley lakes sometimes leave little room for road or railway, as in the middle reaches of Lago di Garda. Ascent from the main valley into the tributary has often been made difficult: it may be made by flights of stone steps, by zig-zag paths or roads with hairpin bends, by ramps along the trough sides, or by wire ropes or helical tunnels.[117] Access may also be had over cols from adjacent hanging valleys.

The deepened valleys and newly created forms have modified, often seriously, the distribution of insolation and its effects on settlements.[118] While in V-shaped valleys cultivation gradually ceases upwards, in U-shaped valleys cultivation and habitation are restricted to the valley floor, to terraces on the sides and to shoulders and the *Schliffbord* above the trough.[119] The *Schliffgrenze* is often the limit of cultivation, e.g. in the Oberalp or Ober Engadin.[120] Rock-terraces flanking the U-valleys carry homesteads, small hamlets and mountain railways, and the *Schliffbord* bears many huts of the Alpine clubs. The steeper parts of the shoulders, often forested, grade into the flat, high mountain meadows of the Alps, the sites of detached communities, of farms and small villages where milk is made into cheese and where wood carving and similar pursuits occupy the inhabitants during the winter months. Corrie floors also are sometimes green meadows.

Rock-barriers, bastions or moraines, exposed to the sun and free from avalanches or floods, have provided numerous sites suited to defence, namely sanctuaries, ancient burgs or castles, and the fortresses of modern times.[121] Barriers or steps with barriers, which are sometimes the sites for dams for reservoirs, as in the Rhine area,[122] may also form political boundaries.[123] They tend to keep the communities above them isolated so that these develop their own characteristics.[124]

The drift is important for water-supplies, for agriculture and for constructional and other purposes (see p. 359). It has influenced the soils and scenery, the routes of roads, railways and canals, the sites of settlements and even the international boundary between Canada and the United States.[125] The lighter and better drained soils of central and north-western Europe, such as loess, sand and gravel, were areas of settlement from neolithic times onwards. Thus the loess in central Europe largely controlled the distribution of neolithic settlements[126] (fig. 98), and in Denmark the early neolithic people preferred the hill-sand to the heath-sand where hard pan discouraged much cultivation in early times.[127] The contrasted drifts of east and west Denmark (see pp. 442, 494) have indeed caused differences in the relief, soils, flora and fauna, and in the crops and types of houses in the two regions[128]: the moraines of the eastern strip are intensively farmed; ancient roads meander through the countryside; and modern roads involved many cuttings and fills. The density of villages and of population is greater than on the outwash of the west, and especially on the morainic clays. The choice of morainic deposits of the *bakke-sand* of Denmark for the Bronze Age burrows is immediately apparent.[129] Roads and railways often cross valleys over moraines, as in the Yorkshire dales,[130] though in north Germany push moraines have jeopardised houses and railways built upon them.[131]

While moraines of strong expression, as in North America, whether interlobate or frontal, are favourable to forest types, the loose, dry sands, sandy loams or gravels of the thicker parts of the outwash sheets, in which the ground-water is low, furnish light soils, as in the "Barrens" of Wisconsin, the heaths of the Alpine glaciation north of the German Alps,[132] or those of north Germany, e.g. the Lüneburger Heide—the bare Baltic Ridge and its outwash were for centuries the frontier between Poland and Pomerania.[133] The sands may have dry deciduous woods of oak and birch, e.g. on the Lower Terrace of the north Alpine zone,[134] and pass into moors where the distal ends of the outwash thin and become finer in texture. Their even surfaces,

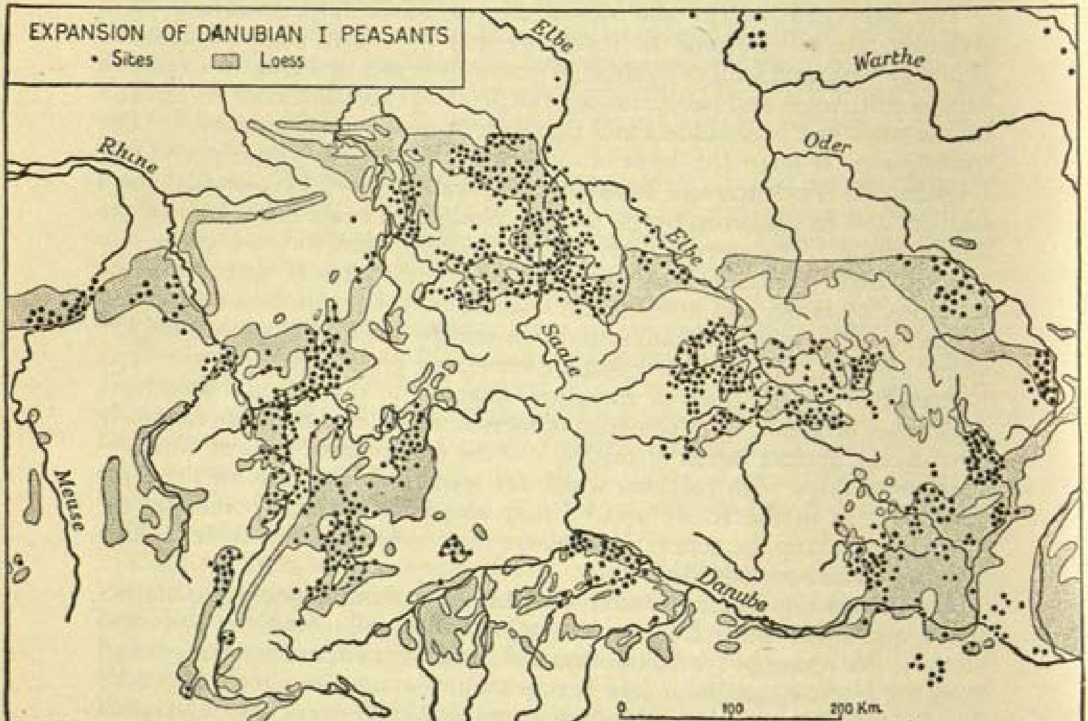


FIG. 98.—Colonisation of the loess of central Europe by neolithic Danubian peasants. J. G. D. Clark, *286*, p. 96, fig. 45.

contrasting with the irregular relief of the moraines where the houses are isolated, permit settlements to grow into hamlets or villages. The gently sloping floors have made it possible, for example, to irrigate parts of western North America with minimum expense.[135]

The Baltic Ridge has strongly influenced the distribution of population and agriculture[136] and the distribution of beech in north-west Germany.[137] Its abundant lakes dictated the strategy of the two battles of Tannenberg (1410, 1914).

Osar embankments, which pass in long windings across vast tracts of country, are also culturally important. Their intersections with water-courses controlled the distribution of early settlements,[138] while they have been used more recently for Indian trails in North America[139] and for paths

and roads as about Uppsala and in *Kultur*-Finland (J. G. Granö distinguished this from *Natur*-Finland) whose towns and villages nestle on their sunny sides[140]; they have even influenced the type of dwelling. Their perfect drainage gives rise to springs and wells at the base,[141] marked by rows of buildings ("dwelling rows"), and causes them as in Finland to be covered with heaths of pine forest. They have provided road material and influenced railway routes in Scandinavia,[142] just as the railway from Leningrad to Helsinki and the towns on Finland's south coast run along the Salpausselkä ridge.

Drumlins control road directions and the division of the land by hedges, and cause houses, either singly or in clusters, to perch upon their drier and sunnier sides.[143] Erratics have been used in monuments, both ancient and modern, and for building stones and road metal (see p. 362).

The beaches of the great glacier-lakes of North America were used as trails by Red man and as highways or "ridge roads" through the forests by the early white settlers, subsequently as sites for cities, for graded tracks for railways,[144] e.g. in Lake Warren and Lake Iroquois, and on the C.P.R. along the north side of Lake Algonquin, the deltas often making excellent sites for stations and sidings.[145] Much of the pioneer settlement and travel was along the beaches, often several hundred feet wide, in the area of Lake Agassiz.[146] The beaches have also become the sites of farms, of summer houses and of villages. In North America, the beaches also gave well-drained ground for vine culture and fruit trees and served as migration routes for many species of plants, including maritime plants[147] (cf. pp. 619, 312). The moraines and beaches of Lake Agassiz have served to site and aid the making of roads,[148] and its floor deposits have provided the fertile soils of Manitoba and structural materials for building and other construction. The floor deposits of pluvial lakes, e.g. Lake Bonneville, have given artesian aquifers.[149]

The coastal deposits of the Ancylus Lake and the Littorina Sea, as in Gotland, are used for roads and settlements,[150] and those of the lateglacial submergence in North America provide sites for villages and roads as well as much of the best farmland.[151]

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Chapter XXV

POSTGLACIAL RE-ADJUSTMENTS

Nature of re-adjustments. The Ice Age interrupted the normal orderly sequence of physiographic development. Hence, at its close the glaciated terrain again became exposed to the old forces which tried to efface its "disharmonic features" as S. Passarage called them (see p. 1063). The forces were especially vigorous lateglacially when, for example, the hillsides were swathed in drift, the British climate was moister (as the shells in the caves indicate[1]), the ground was frozen, and the lower temperature diminished the evaporation[2] and so increased the run-off.[3] The rivers were swollen[4] and showed a higher stream ratio to precipitation than now, e.g. the Oder near Oderberg, the Vistula below Bromberg, the Memel near Ragnit, and in the Alpine foreland the Iller near Altusried and the Isar near Icking.[5] Lateglacially, as also during the Glacial period itself, stream-flow was greater because it was concentrated into a few summer months.[6] Furthermore, melt-waters streamed from the ice and snow; turbid rivers rapidly rasped their channels; frost played havoc on the hillsides; avalanches were active; and solifluxion, wind and water acted freely upon a land swathed in drift and bare of vegetation. Cirque rims were fluted by snowslides[7] and fallen boulders accumulated around the fringes of the aprons of winter avalanche snow at the base of trough walls.

Some terraces, generally regarded as "postglacial", may be lateglacial,[8] as may some of the screes in temperate lands, e.g. in Skye, Wales and parts of the Lake District in Britain[9]; for their accumulation under present conditions is largely checked and fresh fractures or bruises are relatively rare. Some screes in Mull are notched by the 25-ft beach.[10] Slopes were likewise adjusted by landslipping during lateglacial time[11] and the aggradation of the valley floors was partly accomplished while the glaciers were still in retreat.

North of the German Alps, the trumpet valleys in the outwash plains and the associated cones were made during the withdrawal of the ice, postglacial erosion being negligible (see p. 439). The mere recession of the ice caused coarser loads to be laid down nearer and nearer the mountain centres so that lighter loads at a distance led to erosion of earlier deposits.[12] Even in early postglacial time, rivers worked at a quicker pace since rock-barriers, created by ice-erosion or glacial epigenesis, were then undiscovered by the streams: the valleys were in drift and touched rock at few places.

The re-born rivers attempted to re-adjust the valleys to their own needs, in many respects the reverse of those of glaciers. In re-establishing their normal gradients and obliterating the inequalities glaciation had produced, they have been actively transporting the detritus, lodged directly or indirectly by the ice, and recovering from the arrested development resulting from overloading. Glacial floors have been smoothly graded, as in the big eastern valleys of the New Zealand Alps.[13] Lake-bottom deposits have been dissected and valley trains, as in the Mississippi,[14] have had wide swaths cut through them. Convex shapes have given place to concavities and young

narrow valleys, bordered by bluffs of drift (Swed. *Nipa*; Finn. *Mella*) or by river-terraces, have come into existence. Narrow remnants have been preserved along their sides, particularly downstream from projecting spurs, and outwash plains persist only in narrower or wider strips, as in the south German valleys.[15] Broad terraces or narrow shelves record the progress the streams have so far made in adjusting their beds to the present conditions of volume, load and gradient. Streams have regained their erosive power by becoming less heavily loaded and by cutting away barriers of solid rock or drift farther downstream. The Canterbury Plain of New Zealand has been entrenched very deeply—*c.* 180 m in the case of the Rakaia River.[16]

The finer materials have been readily removed, as in the classic earth-pillars of Bozen in Tyrol[17] and the earthpillars of Norway and the Himalayas.[18] Residual boulders bestrew the stream beds[19] and, like the boulder concentrations and pavements (Ger. *Steinpflaster*) along the shores of seas and lakes, arrest the erosive progress, giving rise to rapids, noticeably where morainic barriers, such as the Salpausselkäs, cross their paths.[20] Erosion, relatively slight over the interfluves,[21] has been mainly along stream courses. Channels have been rapidly incised with sometimes an impressive cleaning out of the drift. Clays have been more extensively dissected than sands and gravels which allow of percolation,[22] and eskers and moraines, but little modified, have withstood erosion better than wide plains.[23] Occasionally, as in Württemberg, rains and the prevalent winds have removed the clayey matrix from the windward side of drumlins.[24] Broad glacial terraces, as in the St. Lawrence valley, have been effected by slip and earthflow.[25] The rivers have dropped their detritus as deltas at their mouths: the many local names ending in *ör* show how common these deltas are in west Norway[26] (see p. 479).

The rate of removal of the drift was not everywhere the same. It was higher, for instance, south of the St. Lawrence River than in Labrador[27] and in the western than in the eastern Highlands of Scotland.

Subaerial agencies of postglacial date have hastened to obliterate the harsher features and to restore to the cirques their pristine shape of torrential basins, replacing "alpine scenery" by Mittelgebirge form. Chimneys have been opened along planes of weakness, and walls have been ravined by streams and attacked by frost, as in the *Schuttkaren*, or stripped by slips,[28] as in the summit movements of Rosa Blanchi. Detritus and aggraded waste so brought down hide the break of slope at the base of the cirque walls and commonly conceal the floors. Cirque-lakes, contracted by fans, deltas or screes, are in all stages of infilling, the process being active in the Alps[29] and Pyrenees[30] and well advanced in the Carpathians. Cirques in the soft sandstones of the Apennines have for the most part returned to the preglacial V-form.[31] Dolines have been hollowed out of cirque floors in limestone areas, as in the Limestone Alps,[32] where they may emphasise or diminish the backward slope.

U-valleys are being harmonised with present conditions as in parts of Greenland and Spitsbergen where also cirques have been converted into fluviatile slopes and where roches moutonnées and lakes have been obliterated.[33] Cliffs have slumped to less bold and more stable slopes and the U has been replaced by the V cross-section. Lateral streams have incised shallow clefts in the walls and faceted spurs, and small notches have been cut into the lips of hanging valleys. Where these were open V-shaped, the narrow postglacial ravine has combined with them to produce the Y-valley.[34] Oversteepened

sides in favourable structures have been wasted by slips[35]; slopes have crumbled and been cumbered with talus (some blocks exhibit glacial scratches on one side[36]); and cones and deltas have been heaped up at the mouths of tributaries.[37] Streams have built watershed deltas or "corroms"[38] on the passes to give rise to lakelets, e.g. in the north-west Highlands of Scotland,[39] and elsewhere have segmented overflow valleys, as in the north German *Urstromtäler*,[40] forming lakes upon their floors.[41] Small tributaries have been locally rejuvenated and small postglacial gorges excavated in valleys which glacial melt-waters had overdeepened.[42]

Steps and barriers have gorges (Fr. *gorges de raccordement*; Ger. *Ausgleichungsschlüchte*) sawn across them, usually at their lowest points,[43] or are occasionally pierced by natural tunnels.[44] Screes and fans have accumulated on their downstream side and lake-terraces and deposits of great depth in the temporary lakes above them, as Alpine engineering works have proved.[45] By working back towards the elbows of diverted streams, rivers beheaded by glacial action are restoring their earlier physiography. Stream-action in France has produced the *limons anthropogènes*.[46]

Lake-shores have been eroded and beaches and deltas constructed in their embayments. Drainage and infilling have replaced countless lakes by ill-defined marsh and swamp, lake-flats, and plains as in the *replats* or *Boden* of the Alps,[47] or the grass vales or "parks" of the Rocky Mountains; about half the former lakes of Finland are now peat-bog.[48] Over 100 lakes have disappeared within 100 years in Tyrol where place names like *Seealp*, *Seeberg* and *Seewiesen* are frequent.[49] The lakes in Canton Zürich were reduced in number from 149 to 76 in 60 years[50] and in Schleswig-Holstein from 26·77% of the area to 3·22% (Breckwoldt, 1914). Even rock-basins have been eliminated by alluvial aggradation or by trenching the outlet gorge: exceptions are the pass or col lakes which are little liable to either of these processes.[51] Kettle-hole lakes disappear by growing vegetation rather than by erosion or sedimentation.

The subalpine lakes are filling up so rapidly (see p. 1522) that their future life, measured in years, is estimated as follows[52]: Boden See, 12,500; Thunersee, 13,000; Lac Léman, 20,000 and Vierwaldstättersee, 23,000.

Drift coasts have suffered much, e.g. in Wales,[53] including Anglesey where a strip possibly 2 miles (*c.* 3·5 km) wide has been removed[54]; in England,[55] including Holderness where the coast has been pushed back 2 miles or 3·5 km since Roman times; in north Germany[56]; and in New England[57]—Cape Cod has receded more than one-third of a mile (0·5 km) within historic time.[58] This cliff recession under wave attack has provided detritus for the bouldery shores and boulder-barricades,[59] e.g. on the ends and summits of drumlins,[60] and for beaches and spits, the north German *Nehrung*. It has yielded sands for coastal dunes and determined the character of the adjacent sea-floor, as in the North Sea and the Baltic.[61] Aided by submergence, it has fashioned out of the till watersheds the great ridges of boulders called "sarns" which run out at right angles to the coast off parts of the Welsh coast.[62] In England[63] and along the north German coast[64] accretion of new land has far exceeded the loss.

The phenomenal delta-building habit of the Mississippi is of Quaternary age, due largely to the heavy load derived from the drift of the basin—the margin of the "bird-foot", with the transport of *c.* 350 million tons of sediment, advances seawards 90–104 m/annum. The waves and currents of the

Gulf of Mexico succeeded fairly well in preglacial times in distributing the sediments brought down by the river so that the deposits near its mouth were not much thicker than contemporary deposits to the east and west.[65]

Weathering of drifts and glaciated surfaces. The drifts have been weathered to a depth depending upon their nature and drainage. The finer materials have been removed and the boulders have "grown" and in places protected earth-pillars. The surface on moorlands has been altered to grey, with the removal of iron oxide by peaty acids, and has been oxidised to yellow or brown limonite in the temperate zone or to red laterite in the tropics. Calcareous matter has been leached out and sometimes segregated as "race" or "calcareous rings"[66] at the base of the weathered layer, though it has locally been restored from the neighbouring loess.[67] With the removal of the fine material, the texture has become loose and gritty.

Boulders and pebbles in the more permeable drifts have reacted to the bleaching processes in various ways.[68] They have become soft and decomposed and encased in ferruginous rinds, frequently forming concentric shells of weathered rock, as in basalts, granites and epidiorites, or disintegration pseudomorphs after the original boulders. Limestone boulders have either disappeared, as in south Ireland,[69] or become much fewer; they now possess roughly etched surfaces, their fossils projecting from the face like the porphyritic crystals in igneous rocks. Only quartz, quartzite and chert remain undecomposed. Surface erratics have been faceted by winds[70] or, especially in the case of granites, basalts or gabbros, have had their angles rounded by exfoliation. Laminated or cleaved rocks have been reduced to fragments.

Striae have usually been preserved by a cover of superficial deposits (if these are not too thin or pervious) or along the shores of lakes or seas,[71] e.g. in the uplifted parts of Greenland and Spitsbergen, if lichens are absent, or of banks of rivers below water-level. Elsewhere, the rocks have been roughened, pitted and etched and the striae have been removed in the order of their depth, so that those early discovered are often no longer recognisable. Others have been partially effaced, persisting only as "ghosts of scratches", or have been quite obliterated. The polished skin has scaled off by frost,[72] e.g. in Baffin Land, the high interior of Norway or on the gneiss of west Greenland, or by the action of rain, changes of temperature, or lichens and other growths. The finest striae, however, are still intact on hard, fine-grained and insoluble rocks, such as quartz, on veins in the Lewisian gneiss of Scotland, or as Sefström[73] observed, on the weathered gneiss of Scandinavia.

Glaciated shapes like roches moutonnées, deep flutings and "friction cracks", being more permanent, are preserved far more abundantly, though even they have been modified by the opening of joints.[74] Block weathering has been active on the rugged lee faces of roches moutonnées; in places it has entirely destroyed the glaciated forms.[75]

Changes are not unknown in lower latitudes. Pleistocene laterites, for instance, are being disintegrated and removed by the more frequent rains of to-day.[76]

Minor character of adjustments. Although postglacial erosion along drift coasts and in glacial epigenetic gorges is frequently well advanced,[77] and moraines, e.g. in central Asia, have been buried under later screes or slips[78] and alluvium has accumulated on the Indo-Gangetic plain to a depth of *c.* 210 m,[79] postglacial remodelling is still in its infancy. The decay is often

scarcely appreciable on roches moutonnées, grooves and striae as Sefström[80] noticed. Cirque-lakes have frequently only insignificant notches in their outer rim, and even large streams have been able to cut only extremely youthful trenches in the steps of trough-floors and the lips of hanging valleys. Cirque-cliffs, e.g. the *Spiegelwände* of Zillertal, are often still fresh and the dismantling of the walls of U-valleys has made little progress since talus cones are few and small unless the joints are closely spaced. The severe erosion and rounding the ice accomplished have successfully prevented any large-scale shedding of scree unless the rocks are much foliated or jointed and readily disrupted. Even incoherent and unconsolidated drifts have preserved their glacial aspect. Minor and more delicate topographic forms are scarcely touched; moraines stand out as bold and steep embankments; outwash sheets are but slightly dissected; beaches and lake-shore lines are sharply delineated; and spillways have sharp and unbevelled edges and any streams left in occupancy have carved only miniature gorges. The foot of the Falls of Schaffhausen has receded only 10–30 m.[81] Drumlins retain their perfect form—this has enabled them to resist denudation, since creep has not destroyed their symmetry, and surface-waters have flowed off evenly unless their streams crease the drumlins where these coalesce. Moraines and osar have also their original contours and freshness, notwithstanding an occasional sharpening of their crests or a reduction in the angle of their slope by wash. Lakes abound in undrained hollows in tills, moraines and kettle-holes; generally only the smaller of them have disappeared just as only the weakest barriers have been gashed. Two or more outlets occur in some Scandinavian and North American lakes: the labile drainage is still directed. The flats in U-valleys are only slightly aggraded and deltas at the heads of lakes have grown forward only a fraction of the length of the lake. Upland streams, still in drift, have progressed but little in the task of clearing out their courses, a task rarely accomplished since rock, except in epigenetic gorges, is seldom exposed. Streams are still often unnavigable in countries of low elevation. Rivers in the newer drift, for example of Britain, have had no time to develop a terrace sequence like that of the Thames. Deposition off glaciated coasts has apparently been inconsiderable[82] and the adjustment of the sediments of the continental shelf to the restored sea-level is still in progress.[83]

Periglacial features in Europe, e.g. in Poland and France, have been little disturbed[84] and in pluvial lands, the playa deposits, beds of salt and the shore-lines of extinct lakes, with their cliffs, terraces, beaches, bars and deltas, are perfectly preserved. In both belts the most susceptible to change are the steeper, higher slopes with severe climate and lacking vegetation.

In short, postglacial re-adjustment is very incomplete; the scenery comprises elements of glacial and non-glacial relief and the glacial still largely predominates.

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(*C*) *PERIGLACIAL PROCESSES*

Beyond the glaciers and the ice-sheets and in some measure on nunataks and ice-free areas within their limits, processes are active which though not glacial yet owe their existence and character to the ice. Severe frost, ice-winds and other climatic conditions generate in their turn a certain facies of mechanical weathering and spread the glacial influence by means of drift-ice in its various forms over vast expanses of the ocean. W. v. Lozinski's term[1] periglacial which suitably expresses the zone's relation to the ice has been generally adopted though the boundary with the extraglacial zone is by no means obvious; some of the accumulations described in earlier pages under the rubric extraglacial might also be included here.

The subjoined classification shows the nature of the periglacial processes. In conjunction with that set out for the glacially erosive (see p. 231) and constructive (see p. 361) processes, it completes the classification of the products which arise directly or indirectly from ice-sheets.

Classification of Periglacial Processes and Products

1. On Land
    - (i) Aeolian
        - (*a*) Loess
    - (ii) Cryoturbation
        - (*a*) Permafrost
        - (*b*) Ground ice
        - (*c*) Solifluxion
        - (*d*) Polygonal markings
        - (*e*) Blockfields
        - (*f*) Lake-ramparts
        - (*g*) River-ice
    - (iii) Avalanches
2. On Sea
    - (i) Drift-ice
        - (*a*) Calving floods
        - (*b*) Abrasion
        - (*c*) Glacio-natant boulder-clay

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Chapter XXVI

LOESS

Importance. Loess, by far the most important periglacial accumulation, was first recognised and given its rightful prominence in the valley of the Rhine where it is well developed in the Kölner Bucht and Neuwied basin. Here A. Braun[1] in 1842 accurately described it—the name, which belongs to the language of the peasants and brickworkers of this region, came into scientific literature about twenty years earlier.[2] It was found afterwards elsewhere in Europe and in 1846 along the banks of the Mississippi by C. Lyell[3] and A. Binney[4] who declared it to be analogous with the Rhine loess, though its vast expanse in North America was not appreciated until much later[5] (1873). China's extensive loess was discovered by R. Pumpelly[6] (it constituted his terracic deposits) and described in detail by Richthofen[7] who travelled widely in Mongolia and north China.

Loess, given adequate moisture (in south-west Asia it is often salt or desert steppe), is one of the best agricultural soils of the world,[8] especially for cereals and steppe grasses; the principal wheat producing regions of the world, viz. Ukraine and south Russia, Kansas and other states of North America and Argentina, coincide roughly with its distribution. Some of the best soils of Germany are on the loess.[9] Responsible for the fertility[10] are its porosity and hygroscopicity, its aeration, its lime, alkalis, phosphates and other soluble matter and nutritive salts, its freedom from trees or enclosed stones, its ease of working, and its grain size which allows rain to penetrate readily and yet produces the optimum capillary lift of the water. In Persia and Turkestan the loess has been the basis for independent civilisations. In China, by discouraging a heavy tree growth, it played a prime part in the diffusion of culture; it long furnished sites for settlements, including the most ancient capitals,[11] and was probably the cradle of Chinese culture[12]: thus yellow is the sacred colour. Loess explains the constancy of many European sites during the Palaeolithic,[13] while its freedom from forests, closed or open (trees being unable to root securely in the friable soil), agrees with the absence of trees from the Chinese loess when so much of the country was forested during the 2nd century B.C.[14] On both the *Steppenheidetheorie*[15] and the theory which believes oak forests grew on loess,[16] the distribution in central Europe of neolithic peoples is linked with the loess—the blackearth of central Germany, Bulgaria and Rumania however was free of neolithic settlements.[17] The fact that early man sought such forest-free areas has been shown,[18] for example, for Saxony and Silesia. The loess-covered valley of the Danube provided a highway across central Europe by which the neolithic invaders and Beaker Folk from the Black Sea diffused their civilisation into Bohemia and by the Elbe, Neckar, Main and Rhine into Belgium and north-west Europe[19] (see fig. 98, p. 500). The Slavonic languages may have taken shape on the loess of the Carpathian region and radiated thence eastwards and westwards.

When the postglacial pine forest led to cultural decay and fragmented

groups of people were left on the sea-shore living on shell-fish (see p. 879), the hunting population inhabited the forest-free areas[20] which in this respect resembled the Muschelkalk and sand-dunes of Germany and the Chalk country of north France.[21] The freedom of the loess from trees is shown, for example, by the undamaged mammalian bones and the undisturbed burrows.[22]

The *Steppenheidetheorie* is, however, strongly contested[23] since among other things it may take insufficient account of man's clearance of previous forests. Man's preference for loess soil may be connected with its fertility, its ease of working with primitive implements, its dryness and suitability for pit dwellings and its level surface.[24] Forests may formerly have grown on the loess of the U.S.S.R.[25]

Loess is supremely important too because of the light it sheds upon the climate of the Glacial period and upon the succession of glacial and interglacial epochs and their relation to human cultures and animal life (see chs. XXXVII, XXXIX).

Composition. Unaltered loess, both as superficial soil and mother rock, is physically and chemically distinct from other kinds of rock. Typically, it is a brownish-yellow or buff-coloured, calcareous sandy loam (Chin. *huang-tu*, yellow soil), characterised by a homogeneous structure and a fineness of its constituents which make it so friable that it can be rubbed to an impalpable powder with the fingers. Its materials are undecayed and contain no colloid clays.

But the composition is not so uniform as Richthofen assumed; for later research has revealed considerable variability according to its derivation.[26] Chemical analyses, as for China,[27] the Rhine,[28] Wolhynia,[29] United States[30] and South America,[31] show that its lime ranges from 10% to 25% (German *Bördeloess*, 7·9% to 19·9%[32]), its silica from 60% to 70%, and its aluminium silicates from 10% to 20%. Representative figures are set out in the table below.

CHEMICAL COMPOSITION OF LOESS

	Near Galena, Illinois	*Near Dubuque, Iowa*	*Vicksburg, Mississippi*	*Kansas City, Missouri*	*Peorian loess*	*Villejuif, Paris*	*Kansu, China*
SiO_2 . .	64·61	72·68	60·69	74·46	75·07	59·46	59·30
Al_2O_3 .	10·64	12·03	7·95	12·26	10·21	7·54	11·45
Fe_2O_3 .	2·61	3·53	2·61	3·25	2·24	2·42	2·32
FeO . .	0·51	0·96	0·67	0·12	0·43	0·71	1·55
MnO. .	0·05	0·06	0·12	0·02	0·06	0·07	—
MgO. .	3·69	1·11	4·56	1·12	1·03	0·87	2·19
CaO . .	5·41	1·59	8·96	1·69	1·78	12·25	8·35
Na_2O .	1·35	1·68	1·17	1·43	0·89	1·16	1·80
K_2O . .	2·06	2·13	1·08	1·83	1·87	1·63	2·17
H_2O . .	2·05	2·50	1·14	2·70	4·22	4·06	0·96
TiO_2. .	0·40	0·72	0·52	0·14	0·68	0·78	0·60
P_2O_5. .	0·06	0·23	0·13	0·09	0·29	0·12	0·20
CO_2 . .	6·31	0·39	9·63	0·49	0·62	9·24	8·94
C, organic .	0·13	0·09	0·19	0·12	—	—	—
SO_3 . .	0·11	0·51	0·12	0·06	0·40	—	0·20
Cl . .	0·07	0·01	0·08	0·05	0·06	—	—
	100·06	100·22	99·62	99·83	99·95	100·31	100·03

The mineralogical composition is usually diverse.[33] The sand is principally quartz, to a less extent felspar, hornblende, calcite, pyroxene, rutile, zircon, apatite and tourmaline, while the clay consists of fine mica flakes, frequently with kaolin.[34] The absence of the heavier and darker minerals explains the light colour.

The calcium carbonate which, as in south Germany, is often greater if the bedrock or drift is rich in lime,[35] forms encrustations around the tubes (see below), coats the grains thinly[36] (thereby imparting some cohesion to the loess and preserving the tubular structure), or forms concretions deposited secondarily as a gel.[37] It may have been derived from the underlying rocks by rising ground waters[38] but much more probably from tiny grains originally distributed evenly throughout the mass[39] and from contemporaneous weathering.[40] Chemical changes and the action of capillarity, gravity and oscillations of moisture played their part,[41] the deposition being controlled by the level of the water-table. Occasionally tiny calcite crystals occur or pieces of limestone, e.g. Muschelkalk, contemporaneously entombed.[42]

Much of the lime is below the decalcified layers in definite beds[43] up to 0·5 m thick, or in bizarre nodules, the German *Loessmännchen*, *Loesspuppchen* and *Loesskindeln* or the South American *toscas*,[44] which are frequently the size of a potato or even 30 cm in length.[45] Usually hollow (they are said to be solid in Indiana[46]) and often with loose inclusions (Ger. *Klappersteine*) and deeply fissured in the interior through later contraction, they were formed by percolating waters charged with calcium carbonate and may have grown round a nucleus, such as the droppings of birds.[47] The nodules alone remain if the loess thins on to a hill or has been removed; they build the *Puppensteinfelder*[48] and the tosca plains of the South American pampas.[49]

Surface decalcification and oxidation and the conversion of silicates into clay minerals, which rise in amount from 5 % to 15 %, yield a brownish loam[50] (Fr. *terre à briques*; Ger. *Loesslehm*) or superficial brickearth which is neither true loess nor true loam. The loam zones (Ger. *Leimenzonen*), which in places divide the loess into separate horizons and provide separate water-tables, may be only locally significant and dependent upon drainage.[51] Alternatively and more probably, they have stratigraphical importance and indicate a moister climate which permitted humus and carbonic acids to form from decaying vegetation[52] (of which traces remain). This is proved by the occurrence, as in Austria, of molluscs indicative of warmer and moister conditions.[53] Thus the three loess horizons recognised in Austria are correlated as follows[54]: the Krems zone with the Mindel-Riss, the Göttweig-Wielandsthal zone with the Riss-Würm, and the Paudorf zone with the Würm 1–2. In south Russia and the Ukraine too the loam zones contain fossil black-earths which pass laterally into chocolate-coloured steppe soils or podsols.[55]

Size of grain. The grains which are sometimes round or subangular[56] but usually sharp and unattacked by weathering, as Richthofen[57] affirmed for China and Wahnschaffe[58] for north Germany, are extremely fine, as A. Jentzsch[59] first demonstrated. Mechanical analyses,[60] perhaps not irreproachable in their methods, reveal a vast preponderance of grains 0·05–0·01 mm in diameter, with a nearly symmetrically decreasing series of admixtures above and below (fig. 99). Some loess is still finer, as in Russia and North America where respectively 97·3 % and 97·7 % of the grains are less than 0·005 mm.[61]

R. Ganssen,[62] who fully discussed the chemistry of the loess formation but

in a manner perhaps not wholly satisfactory, ascribed the uniform composition and grain size to secondary removal of the alkalis, a nearly absolute hydration of the fine-grained aluminium silicates and a coating of the fine dust with calcium carbonate. Others attribute the uniform grading to the mechanical disintegration (insolation and frost) that gave rise to the loess[63] (see p. 523).

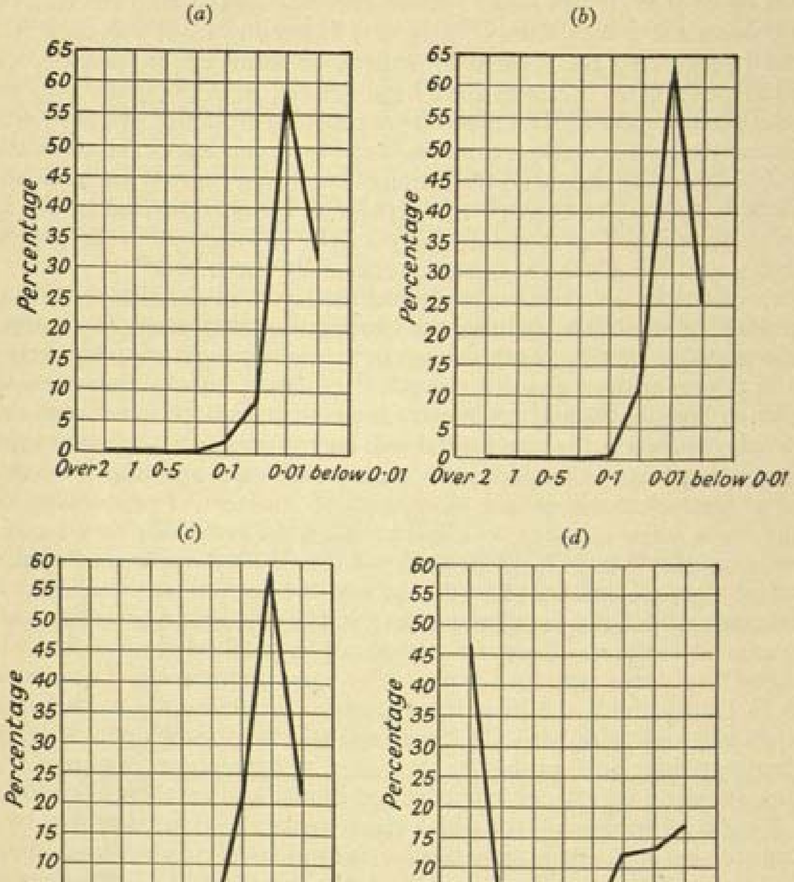


FIG. 99.—(*a*) Loess from Hecklingen (Stassfurt).
(*b*) Loess from Hecklingen, after treatment with acetic acid.
(*c*) Blackearth from Körbisdorf (Halle).
(*d*) Base of the Pariser loess, Ehringsdorf.
Grain size in millimetres.

L. Siegert, *Z. D. G. G.* 64, 1912, p. 517B.

Structure. The uncompacted nature, which won the name for the deposit in the German Rhine valley (see below), is due to processes that are not perfectly understood,[64] but mainly, as Richthofen suggested, to the presence of innumerable, calcareously lined, root-like tubes, of a diameter of 0·2 mm or more.[65] This tubular structure explains the high permeability and porosity of the loess (especially vertically), which in the Ukraine, for

example, is 42·46% to 46·48%,[66] and in the Danube 40% to 47%[67] and may be as much as 65%,[68] and promotes effective drainage and leaching at higher levels and favours the deep penetration of roots. It has made possible the construction of cellars for the storage of wine and potatoes in Austria and the Rhine country and of the cave-dwellings of Sandomierz,[69] Crimea and Dobrudja[70] and of China[71] where whole villages are half-dug, half-built into the cliffs and hills—chimneys and air-vents from the underground rooms present an unusual sight in the tilled fields above. It is also responsible for the infiltration of surface waters which may emerge at the loam horizons.

When dissected owing to a revival of erosion, the weak, pulverent loess, in the absence of surface run-off, rarely slips or crumbles away like the residual clays it superficially resembles, but breaks off in vertical or sharp-faced cliffs along the sides of ravines and canyons[72] (Russ. *owrag*) and in crevasses, pinnacles and natural arches.[73] These faces are also seen in the depressed roads or "hollow ways"[74] (Ger. *Hohlwege*), developed by traffic, wind and rain storms; they may be 30 m deep in Asia and South America but occur to less depths in North America and the Rhineland. This characteristic, which may also be due to tension or shrinkage,[75] is implied in the word *loess* which comes from *loesen* to detach[76] and in the term the "Bluff formation", which G. Swallow[77] introduced into North American glacial literature (pl. XXIA and B, facing p. 544).

The basal drainage, described by Richthofen[78] from China, may excavate hollows which, when their roofs collapse, create surface depressions ("loess wells", e.g. the Wallachian *crovuri* and some *dolines* in the Banat). Such horizontal sapping also leads to a vertical chimneying (pl. XXII, facing p. 545).

The surface of the south Russian loessic area is studded with countless peculiar, drainless hollows of various sizes. These "pods" probably owe their origin to irregular deposition of the wind-blown dust[79] on a basement which was itself irregular, the hollows later being deepened by the solution of the salts in the loess, though some authors have regarded the pods as dolines.

The homogeneity causes the loess to be unstratified, though differences of the layers in chemical composition or in fineness of grain may initiate terraces. Sometimes the loess encloses nests, pockets or strips of loam, sand or clay, owing to varying wind-velocity or to sorting by water in the case of lake-loess (see below). Stratification may be imparted too by a layered arrangement[80] of shells, concretions, boulders, pebbles or gravels. Thick beds of sand, frequently current-bedded, are apt to occur towards the bottom; they constitute the *Sandloess* and some of the "valley loess" (Ger. *Talloess*) of the Rhine and G. Steinmann's *Recurrenzzonen*.[81] Rolled nodules and pieces of older loess are also there. A pebbly base, the *Steinsohle* of German geologists, is common in north Germany,[82] especially near the Mittelgebirge (fig. 100), and is general in the Mississippi and Missouri basins[83] where it may be 10–15 ft (3–4·5 m) thick.

Distribution. Loess has a two-fold distribution: (1) periglacial garlands of "cold" loess[84] about the Pleistocene ice-sheets, as in north Europe from the Ural Mountains to the Seine, around the Alps on the north, west and and south-west, extending into Lower Bavaria and Lower Austria and in a narrow zone north of the Pyrenees; and in North America and South America in similar conditions; and (2) "continental"[85] or "warm" loess in the inland basins and steppes encircling the modern deserts of central Asia[86] between 52 and 56° N. Lat. from north China and the slopes of Tienshan, Alai,

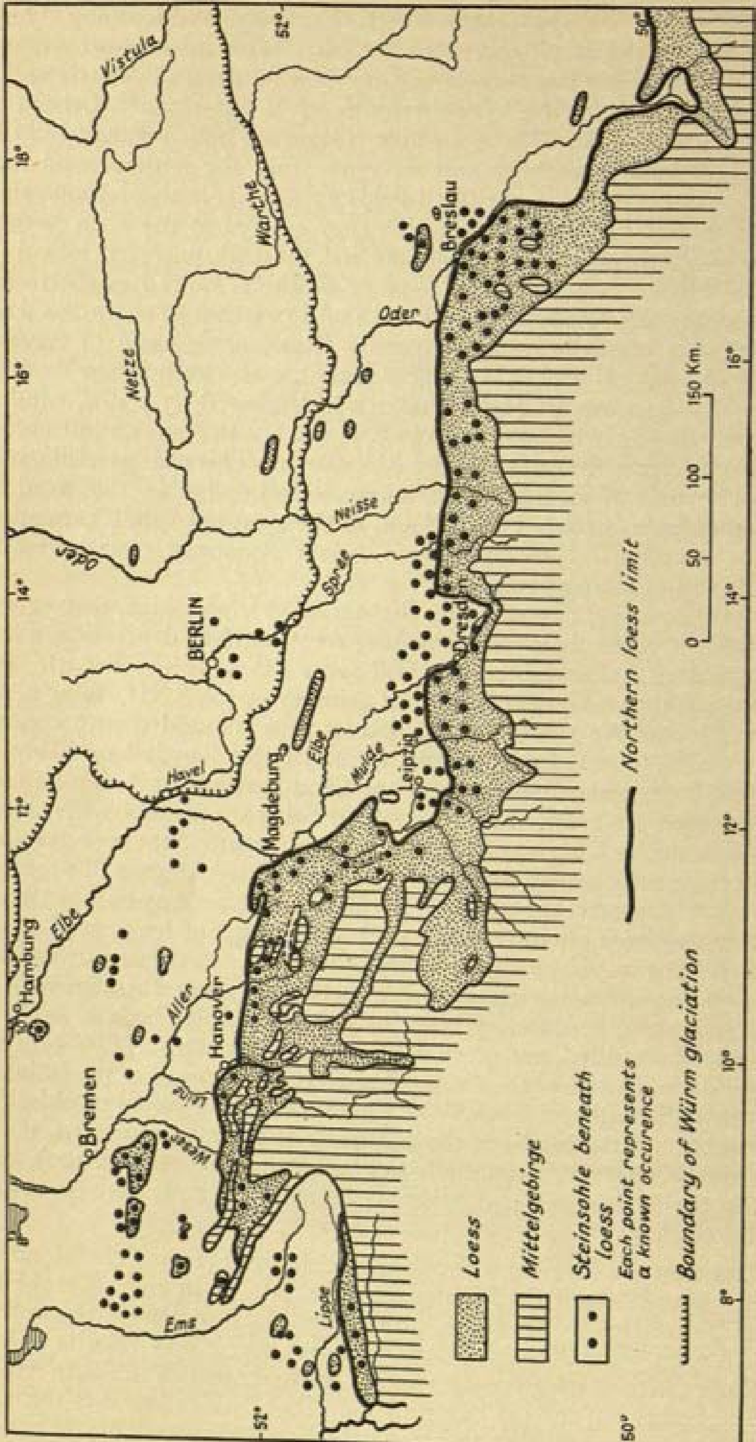


FIG. 100.—The distribution of the loess and the *Steinsohle* beneath the loess and drifted sand in central Europe. H. Poser, 1908, p. 36, fig. 3.

Pamirs, north-west India, Persia, Turkestan, Tibet, Transcaspia (almost completely surrounding the Aralo-Caspian area), Asia Minor to Transbaikalia.

Keilhack[87] who mapped its distribution estimated the area (on somewhat insecure data) to be Eurasia 16 million sq. km, North America 5 million sq. km and South America 5 million sq. km. Of the total, approximately half averages 10 m thick. Walther[88] estimated the volume at 40,000 cu. km, Keilhack at 130,000 cu. km.

1. *Periglacial Loess*

Distribution. The periglacial loess of the Scandinavian ice-sheet (fig. 101) crosses north France in a band of varying width, occurring as a broad strip near the coast in north Brittany[89] and in the basins of the Loire and Seine (cf. map of French loess[90])—the identity of the French with the Rhenish loess was recognised as early as 1838.[91] After passing across Belgium[92] (Rutot's *Hesbayen*) into the lower depressions of the Ardennes it continues as a narrow belt on the North German Plain,[93] usually not more than 30 km out from the Mittelgebirge and generally to a line which parallels these mountains (see p. 533). It sweeps over the older river-terraces, as in Thuringia, on to the northern flanks of the Mittelgebirge, e.g. the Harz, Erzgebirge and Riesengebirge, and as "plateau loess"[94] on to the inter-stream watersheds. In north France it rises to 200 m; south of Aachen to 300 m; in the Oder and Vistula basins to 400 m; on the northern slopes of the Sudetes to 500 m or even 580 m[95], gaining access to Bohemia through passes at 520 m and invading Moravia, e.g. Předmost, by the Moravian Gate; in the Swabian Alb to 900 m; and in the Carpathians to 1200 m (Range, 1955).

A facies of the loess is the *Flottlehme* and *Flottsande*, a mixture of loess and drift sand which covers considerable areas in the Fläming and Lüneburger Heide and occurs as "islands" in north-west Germany.[96]

The northern limit,[97] where the loess thins out, runs eastwards as a sharply defined wavy line from Sangatte and Dunkirk by Courtrai, Oudenarde, Louvain and Köln (Holland has no true loess north of Limburg) and continues generally along the line of the Elbe by Paderborn, Minden, Braunschweig, Magdeburg, Bitterfeld, Leipzig, Meissen, Dresden, Frossenhain, Görlitz, Liegnitz, Breslau, Trappau (on the Oder) to Jósefów (on the Vistula). In the Low Countries, Rhineland and Russia, the loess lies without the drift but over the intervening stretch the distributions partially coincide.[98]

The southern limit is much less easy to define since the loess here thins out and resembles the adjoining residuary clays and loams. In the wetter west, it breaks up into weathered and decalcified "islands".

In Europe (unlike Asia), it is seldom thick enough to develop its own topography though it masks the gullies and depressions where the pre-loessic relief was mature and thins out against the higher ridges.[99] In many places, it has displaced the smaller streams from their old courses.[100] The buried relief and post-loessic denudation explain its varying thickness which rarely exceeds 8–12 m though 35 or even 60 m have been recorded,[101] the greatest occurring in the valleys, e.g. in the upper and lower Rhine (Düsseldorf), in Saxony, the Elbe valley of north Bohemia, and locally in the Danube valley, lower Bavaria and near Vienna.[102] Representative figures[103] are Brabant, 5–10 m; province Luttich, 10–15 m; Rhineland, less than 10 m; Thuringia, 1–4 m; Silesia, 1–2 m; and west Europe generally 3–5 m.

From Germany, the loess spreads out over the plains of Poland, south of the Middle Polish moraine,[104] e.g. about Kraków and in the sub-Carpathian zone, and broadens into Russia[105] where, banked against the slopes of hills and spread out over plateaux and plains, it finds its maximum European development. South of 56° N. Lat. (cf. B. A. Keller's map[106]), its mantle, 700 miles (*c.* 1200 km) broad and 500,000 sq. miles (1,300,000 sq. km) in extent, stretches as far as the Black Sea and the north shore of the Caspian and into Circaucasia[107] and eastwards to the enlarged Caspian Sea in the middle and lower Vistula (see pp. 527, 1131). It covers much of the Dobrudja,[108] where it is up to 80 m thick, and a zone in the Danubian plain of north Bulgaria about 20–30 km broad.

Thinning east of the Dnieper, the loess reappears in Siberia[109] where, though as yet not well known, it covers half of the Kirghiz steppes, the foot hills of the Altai, the slopes of Tienshan and Pamir-Alai ranges and the Fergana and Ili valleys, and is encountered in the eastern half of the Minusinsk basin, near Irjutsk, in the southern zone of Transbaikalia, and in the north between Nijni Vilui and Lena rivers and westward and eastward from Yakutsk. Towards the north in European Russia, it sends out tongue-like penetrations toward Vitebsk and Moscow and along the right bank of the Volga to Kazan.

The south Russian loess averages 16 m[110] (the earliest figures[111] fall within this) but decreases eastwards. Much of it, originally yellow in colour, has since acquired a top stratum rich in humus, the product of a change in climate. This blackearth or *Tschernosem* also thins eastwards, e.g. from 140 cm to 50–60 cm, though it tends to thicken if the mother rock is coarse grained and to be exceptionally thin on limestones.[112] The dark colour, ascribed to marine muds largely derived from the disintegration of black Jurassic shale,[113] to inundation by the waters of the Caspian and Black Sea[114] or to swamps,[115] is due to humus obtained from peat[116] or trees (affirmed by early Russian geologists[117] and by O. v. Linstow for the German blackearth) or more probably from steppe vegetation[118]—there is no sign of a moister climate or of the removal of calcium carbonate or alkalis, and the Russian blackearth has had no forest cover since it was formed during the last glaciation.[119] The thickness, chemical composition and relationship to climate support this view which has been specially elaborated by V.V. Dokutchaiew.[120]

Analyses[121] prove that the humus varies regionally; its maximum is 16% to 18% east of the Volga. Dokutchaiew's map[122] shows by "isohumic bands" how the percentage decreases from its average of 6% towards the north-west (here it passes into grey forest ground and podsol with the increasing precipitation) and south-eastwards where, with increasing dryness, it grades into the greyish-brown soils of the dry steppes. In all cases, it diminishes in depth, finally ceasing very abruptly.[123]

Blackearth also occurs in drier areas, about Belgrade,[124] in Wolhynia,[125] near Vienna[126] and in many German localities,[127] especially in the east and between Magdeburg and Leipzig—the depth in places is 60–80 cm. The total area in Europe comprises 1·6 million sq. km of which 1·42 million sq. km are in Russia.[128] Blackearth is also found outside Europe in China,[129] Argentina[130] and in the United States[131] where it covers 1·7 million sq. km and occurs especially in Nebraska, Arkansas and Dakota.

Modern soil studies[132] show on the one hand that blackearth may descend from solid rock of various kinds and from glacial deposits other than loess, and on the other hand that different soils may be linked with loess according to the

climate. Loess is in this respect important only because its porosity and lime enable the blackearth to extend farther in humid districts.[133] *Tschernosem* is formed under weak alkali or weak acid reaction and is favoured by extreme cold and long winter frosts and by high temperatures and high saturation coefficient during summer, all of which result in high evaporation and desiccation of the soil; winter frost and summer drought alike retard the decay of humus.[134] It has been degraded postglacially.

Loess is associated with the central European glaciations, notably the Alpine.[135] It is well developed in the bigger Alpine valleys and along the whole northern edge of the range from Vienna to Lyon. It occurs in four main areas: (*a*) between Basle and Schaffhausen and south to Aarau—at Basel it is 20 m, at Aarau 10 m thick; (*b*) in the Thur valley; (*c*) in the Rhine valley from Bonaduz via Chur to Feldkirch; and (*d*) in the Rhône between Lac Léman and Leuk. It extends along the Rhine, the classic area, e.g. between Basle and Mainz and on the heights of Kaiserstuhl (350 m), Siebengebirge and the Eifel, in the Neckar, Main, Moselle and Meuse, and spreads along the Danube[136] to the Balkans.[137] It is widespread in Bavaria (Franconia) and the Swabian stepland which it climbs to 900 m.[138]

In France[139] the loess or *limon des plateaux* forms one of the four terms of the Diluvial succession: the others were the *Diluvium rouge*, the *Sable calcaires avec coquilles* and the *Diluvium gris*.[140]

On the polar side of the ice-sheets loess is virtually unknown though it has been alleged that loess occurs in north-west Norway[141] and in the coastal belt around the Arctic Ocean,[142] including Alaska (see p. 528) where, although it is thin and masked by or assimilated in later sediments, it underlies a considerable area.

The loess of the United States[143] covers tens of thousands of square miles outside the drift, but unlike the European loess, is frequently interbedded with the drift over considerable areas. It skirts the Driftless Area (see p. 727) on the west, underlies most of Nebraska and Illinois and much of Iowa and the prairies between Missouri and Texas. The principal distribution is along the Mississippi, Missouri and Ohio (fig. 102) and the average thickness is perhaps about 2–3 m,[144] though higher figures, ranging up to 55 m., have frequently been mentioned.[145] In central Nebraska, the general thickness is not far from 30 m.[146] Thinness and extreme irregularity probably explain why few distributional maps have been published. The eastern bluff of the Mississippi from the Ohio River to the Gulf is mantled with loess to a depth in places of 100 ft (30 m) into which the small streams have cut numerous narrow and deep gullies. The loess here completely dominates the landscape, its bluff rising 125–250 ft (38–76 m) above the flood plain except at the south end where it is lower. Farther east, it changes into a brown or yellow loam to a maximum distance of about 100 miles (*c.* 160 km).

Farther east, the loess becomes patchy and thin: in Indiana and Ohio it is not readily mapped in the field and still farther east is only locally recognisable, e.g. in the Connecticut valley and about Boston.[147]

In Iowa, the most representative loess state, the loess areas are customarily treated under three heads. In the first, the loess, being of uniform thickness, has no influence on the relief; in the second, it accentuates or otherwise modifies the topography; and in the third, it has produced a surface which is largely constructional. In the south-east of this state and in north-west Illinois the loess forms mounds or "pahas".[148] These low, parallel

FIG. 102.—Map of the North American loess (black). A. K. Lobeck, *1921*, pl. 394.

and elliptical ridges or domes, up to 10 ft (3 m) high and 1 mile (1·6 km) long, form a belt 5–10 miles (8–16 km) broad. They consist of loess or of Kansan drift or bedrock capped by loess, the cores being either sand-dunes built up by easterly ice-winds[149] or fluvioglacial sediments channelled by streams.[150]

The Peorian loess,[151] which is up to 60 m thick and is found in Nebraska, Illinois, Minnesota, Missouri and Iowa, extends from near the eastern border of the Des Moines Lobe some 60 miles (96 km) to the Mississippi and beyond into the Driftless Area for a distance of at least 10 miles (16 km) where it thins out and becomes interbedded with surface soils. It spreads down the Mississippi to the Gulf of Mexico.

Periglacial loess underlies the South American pampas[152] through 20° of latitude (20° S. to 40° or 42° S. Lat.) from Bolivia to Patagonia (fig. 103). The resemblance to the Rhine loess is strong but the material is much finer[153] and has less lime and quartz.[154] The depth is usually 10–30 m but is occasionally up to 298 m as proved by bores.[155] Its Pleistocene age is now generally admitted,[156] though the *Postpampeano* or upperost beds are Holocene and the lowest, the *Prepampeano*, are Pliocene.[157] The Pampean contained a mammalian assemblage which included huge sabre-tooth cats, short-faced bears, giant racoons, chinchillas, large hoofed animals, and a wide variety of edentates: its age is roughly that of the Equus Beds of North America (see p. 1260).

Periglacial loess occurs in Hawaii[158] and in the eastern plains of New Zealand,[159] e.g. in Banks Peninsula, about Oamaru and Timaru and in Southland (Willett, 1951, map). Loess of earlier geological periods has been discovered in several parts of the world.[160]

Pleistocene age. The periglacial loess has provoked more controversy than any other Pleistocene deposit. Strong disagreement about its source, mode of deposition and exact age may indeed suggest that its origin was not everywhere precisely the same: relief, drainage, climate and vegetation may have introduced variety. Nevertheless, it has long been admitted that the European and North American loess is Pleistocene and in some way concerns the ice-sheets. É. de Beaumont[161] noticed that the loams were on the equatorial side of the drift. Others[162] found that the loess in Russia and in Europe generally lay outside the erratic limit, though detailed investigations[163] have subsequently shown that the relationship, like the alleged exclusion of loess by the Upper Diluvium in north Germany,[164] is not exact. It is, however, sufficiently accurate to betray a genetic connexion between loess and glaciation. This has been established too for South America. South of *c.* 44° S. Lat. lie the moraines and outwash; then the sands begin, as S. Roth proved; and about the Rio Negro in 40° S. Lat. true loess appears and can be followed beyond the Tropic of Capricorn.

Source. The source of the loess is less certain. L. S. Berg,[165] in his eluvial "soil hypothesis" (*Bodentheorie*), thought it resulted in a dry climate from deep autochtonous weathering of various kinds of rock, an opinion which had been previously expressed by various workers.[166] This is apparently true of the adobe of arid North America[167] and has been applied to the true loess of the lower Mississippi valley which, it is said,[168] was formed by weathering and colluvial transportation from terraces and from the mantle conforming to the topography, especially in dissected terraces—the jointing

is due to tension associated with creep and the tubular structure to coating around living plants. Nevertheless, this view is inapplicable to loess which has fresh minerals, a great thickness, fossil soils, and a sharp physical contact with the rock beneath.[169] Loess is also referred to the action of wind on weathered rocks,[170] e.g. Eocene and Oligocene sands in the Paris Basin[171]

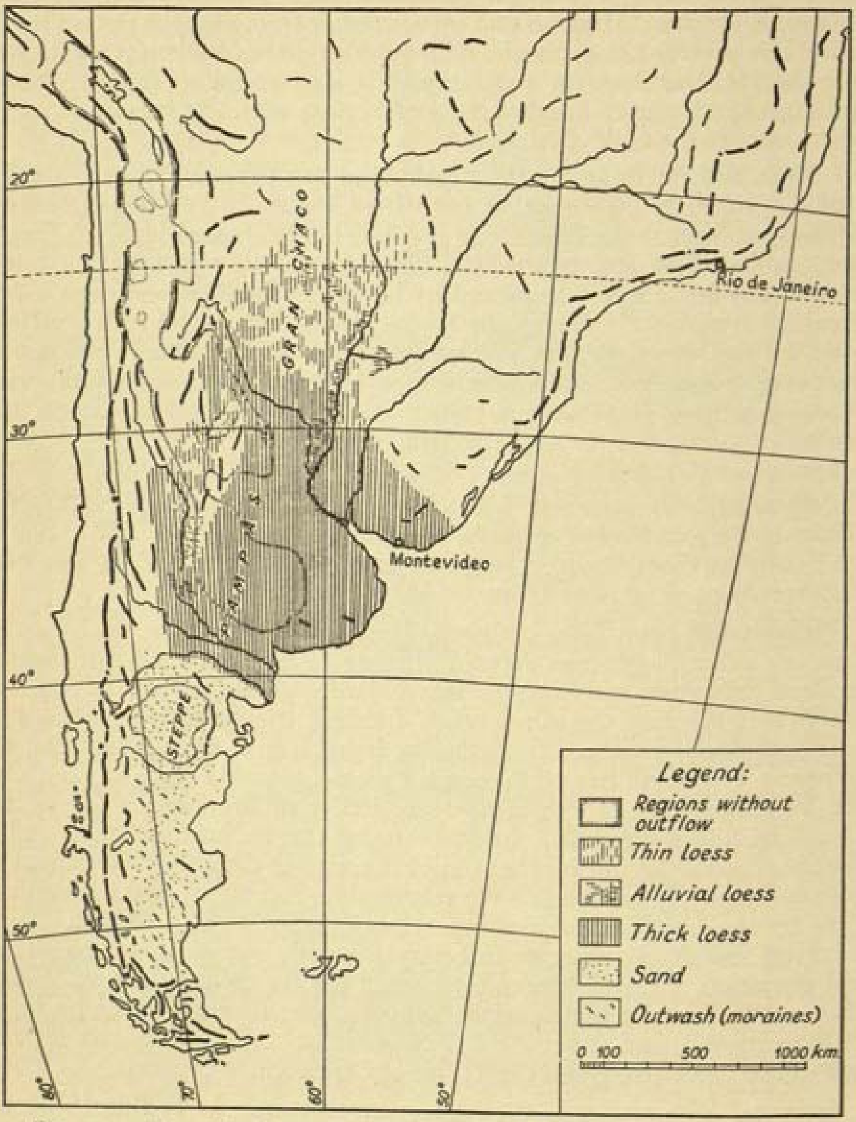


FIG. 103.—Map of the South American loess. A. Scheidig, *1494*, p. 23, fig. 14.

(as implied by the rounded grains of quartz, glauconite and brookite), weathered Flysch in the Vienna Basin,[172] Tertiary sands in Moravia,[173] Keuper Marls at Nuremberg, Lower Cretaceous on the borders of the Teutoburger Wald, and Tertiary and Senonian deposits in south Limburg[174] (proved mineralogically). Alternatively, it resulted from vulcanicity[175] or

from the leaching or alteration of Pleistocene flood-plain silts and clays, as in the valley of the Mississippi[176] (which was then a braided river[177] and not a meandering river as it later became), or from the dust that the freeze-and-thaw process produced,[178] e.g. in forming the *Felsenmeer*[179] (see p. 574) or the frozen ground and frost soils[180] with their pipkrake (*ohne Frost kein Löss*), solifluxion and polygonal ground[181]; silt is the final product of frost weathering in a cold climate with little chemical action. But all these, except the last, which was probably a factor, must be rejected since the loess is exotic and independent of local rocks: in Kansas, the loess is of uniform composition over 30,000–40,000 sq. miles (78,000–107,000 sq. km) and rests on Pennsylvanian, Permian, Cretaceous, Pliocene and Pleistocene accumulations.[182]

Loess is also related to river-muds in the arid areas of the Andes,[183] to the washing of till or moraine by glacial flood-waters,[184] or to winnowing by wind,[185] since its lime content approximates to that of till and the loess frequently rests upon large erratics, the residue of denuded *moraine profonde*. Wind removed the finest fractions from muds exposed on the dry floor of the English Channel,[186] from superglacial dirt on the margin of the ice[187] (south Russian loess becomes finer, less quartzose and richer in Al_2O_3 and Fe_2O_3 as the distance from the ice increases[188]), from glacial sands like those in the German *Urstromtäler*,[189] or bedded loess on river-terraces,[190] schotter plains,[191] or flood plains in process of construction.[192]

A derivation from glacial muds in streams which issued from the ice (when the land was lower and the drainage more sluggish[193]) is widely believed.[194] Thus the loess is thickest along valleys leading away from the ice, e.g. the Rhône and Garonne rivers, and in Siberia, and is coarsest and least clayey near the ice-edge[195]; its iceward border is abrupt; its silicates and carbonates of calcium and magnesium and its particles of felspar, amphibole and pyroxene are fresh; its grain in Lower Austria resembles that of the muds from the Schwarzensteinkees and Morteratsch Glacier[196]; and it is absent east of the Alps[197] where there were no flood plains.

While the loess may not have sprung from the higher parts of the outwash plains whose rapid streams carried it away, flood plains when sun-dried and winnowed of their finest dust or rock-flour, a process not infrequently seen in arctic lands to-day (see below) and on the outwash of the Rhine (see p. 528), would certainly provide, by their constant replenishment, an inexhaustible source for just such an accumulation in respect both to its wide extent and its want of bedding. This view is sustained by the following: the loess occurs in the lee of outwash plains[198] and reflects their characters in both composition and colour[199]; its thickness and grain-size diminish exponentially away from the source of supply, as in the Elbe and the Mississippi valleys[200]; and flood plains exist without fine clays but with dunes.[201] Flood plains, like those of the Seine and Scheldt and the rivers of the Sudetes and Carpathians, which were fed in spring from melting snows on high ground, may have contributed.[202]

The principal sources of the North American loess were the interlobate moraines and outwash, the Iowan drift—mineral analyses of the Loveland Loess correspond closely with those of the Kansan till[203]—and the broad flood plains,[204] e.g. Missouri, Illinois, Wabash, Ohio and Mississippi. Even if the deserts of the west contributed (see below), it should be borne in mind that the Platte and Missouri rivers and all their Rocky Mountain tributaries were in the Pleistocene overloaded, glacially fed streams.[205]

It is occasionally objected, on quantitative grounds, that the loess could not

have been derived from ice-sheets or their drifts.[206] Accordingly, the loess is referred in Europe to the Sahara Desert[207] or the inland basins of central Asia,[208] and in North America to the dry Mesozoic and Tertiary plains east of the Rockies[209] (Colorado, Wyoming), especially since red dust has recently been blown afar from this and other deserts[210] and loess, at least in east Nebraska and Kansas, grades into sand and this in turn was derived from solid rock. Loess is still forming in South America, its production depending upon the strong aggradation on the extremely arid slopes east of the Andes.[211] Others seek its source in volcanic eruptions,[212] e.g. in South America and Europe—modern Hawaii suggests this—or in some cosmic or extratelluric origin,[213] e.g. volcanic activity of the moon, a view akin to that which regards loess as cryoconite[214] laid down at the melting of the ice.

Fluviatile hypothesis. Of the many hypotheses about the loess, those older ones which regarded it as a glacial clay[215] or invoked the sea,[216] estuaries[217] or great diluvial floods,[218] are now only of historical interest. Deposition in lakes, ponded either by a sinking of the land[219] or by morainic or ice-barriers,[220] especially in the Rhine valley[221]—the lake rose to *c.* 500 m and exterminated the mammoth, woolly rhinoceros and palaeolithic man in Europe[222]—and occasionally in the valley of the Po,[223] has found some favour even recently. It is, however, inconsistent with the great altitude of the loess (see above) and with the high lime content, terrestrial shells, and lack of stratification of the loess.

A fluviatile origin has won much approval,[224] notably among modern Russian geologists.[225] Many facts, it is said, are favourable: the thickest and most porous loess lies along the valleys, e.g. Rhine, Danube, Missouri and Mississippi; the lowest layers are indisputably water-laid[226] (the lack of stratification in the loess proper is explained, on this hypothesis, by the fine and homogeneous structure[227] or by the subsequent action of plants and animals[228]); the loess passes into water-borne sands[229] and from loose, open deposits into loamy clay[230] and is occasionally interstratified with boulder-clay[231]; its stones are too big for wind transport[232]; it has bedding, ripple-marks and shells in pockets[233]; and its delineation with the rock-foundation is sharp.[234]

This origin is undeniable for the stratified silt in the loess which is a local pond facies and for the *Steinsohle* and basal sand-loess which in Germany, Russia and North America encloses fresh-water molluscs—that the loess itself generally has none of these shells is attributed to the coldness and siltiness of the waters.[235] Yet it can hardly apply to the bulk of the loess with its terrestrial molluscs (see below) and peculiar physical and topographical characters—attempts have been made to discriminate between water-laid and wind-borne loess.[236]

The molluscan evidence is decisive upon this point. Loess contains countless individuals but comparatively few species, all of them it is said still living species.[237] European lists have been compiled by Soergel[238] and North American lists by Leverett.[239] A. Braun[240] collected over 200,000 snail shells from the Rhine loess between Basle and Bonn and found only 33 fresh-water shells, the rest being almost all *Pupa* [*Pupilla*] *muscorum*, *Succinea oblonga* [*arenaria*] and *Helix* (*Fruticicola*) *hispida* (fig. 104), a uniformity later established[241] for the Rhône and Swabia and for Europe generally. D. Geyer[242] affirmed that these species are found in other glacial deposits and that *Hygromia montana suberecta*, *Xerophila* [*Helicella*] *striata*, *Nilssonia* and

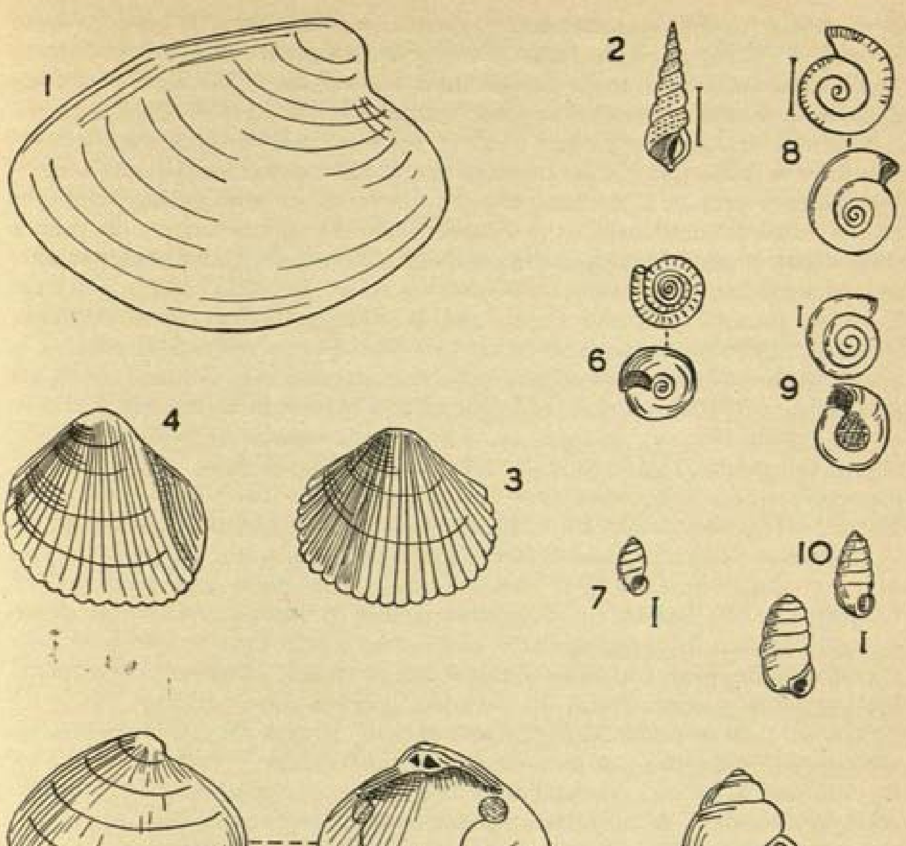


FIG. 104.—1, *Tapes senescens* (*T. aureus* var. *eemiensis*); 2, *Bittium reticulatum*; 3 and 4, *Cardium edule*; 5, *Mactra solida*; 6, *Helix hispida*; 7, *Pupa muscorum*; 8, *Planorbis gredleri*; 9, *Vallonia tenuilabris*; 10, *Columella columnella*; 11, *Viviparus fasciatus*; 12, *V. diluvianus*; 13, *V. glacialis*; 14, *Valvata goldfussiana*; 15, *Lithoglyphus naticoides*; 16, *Corbicula fluminalis*; 17, *Pisidium asteroides*. Natural size except where measuring line is given. F. J. Faber, *477*, p. 394, pl. xvii.

Sphyradium [*Columella*] *edentulum columnella* are more typical. *Limnaea ovata* [*linosa*], *Clausilia parvula*, *Pupa* [*Columnella*] *edentula*, *Hygromia montana*, *Vallonia costata* and *Vertigo parcedentata* also occur. The successive loess deposits in Kansas have characteristic faunas.[243]

Mollusca are more abundant in the older loess and in the lower levels[244] (possibly because water was absorbed less readily as the loess thickened[245]) and are not arranged mechanically[246]: shells commonly exceed 5000 per cu. ft (*c.* 28,400 cu. cm) in the Peorian loess.[247] They are mainly fragile shells (Charpentier[248] noticed this) which have not been transported, and are of terrestrial habit, as Gumbel[249] early showed for Bavaria, T. Belt[250] for the Danube, Braun[251] for the Rhine and B. Shimek[252] for North America. Thus 43 of the 49 molluscs in Illinois are land forms, the majority living in open woodlands or among shrubs far above streams.[253] Aquatic shells are insignificant in both number and species and belong to forms which live in streamlets that dry up in summer. They are common in bedded loess,[254] e.g. in the plastic, thinly bedded "lake-loess"[255] or in some "valley-loess", especially where tributaries enter the Rhine from the Vosges and Black Forest. They are embedded in loess which rains or melting snows have re-deposited as "dejective loess" (Ger. *Schwemmloess*), particularly at the foot of steep slopes or cliffs.[256] This marginal facies or "slope loess" (Ger. *Gehängeloess*[257]) is usually wedge- or cone-shaped, has no capillary structure, contains broken or rolled nodules and is frequently impure. It is thinner than the typical loess and more compact and resistant to pressure. Its coarse, local material is angular near the rock and exhibits slope bedding.

The deluvial or proluvial hypothesis of A. P. Pavlow[258] and other Russian adherents[259] supposes that loess consists of the products of rock-decay washed by rain and deposited on declivities and plains. It resembles the *ruissellement* hypothesis of A. de Lapparent and other French geologists (see below) and encounters many of the objections raised against the general fluviatile theory, viz. the occurrence of loess on water-partings and its uniform composition and want of bedding.

Aeolian theory. It has been widely held that loess was formed by glacier-winds laden possibly with snow[260] (melting snow may have produced the infrequent bedded loess[261]) and grew possibly very slowly so as not to smother the living shells[262] yet, especially near the margin of the ice-sheet, sufficiently rapidly (a fraction of a millimetre per annum has been mentioned[263]) since the shells and bones and the loess itself are unweathered[264] and the subloessic slopes are free from scree.[265] The variable grain-size and sparse vertical distribution of fossil snails of the Iowan loess are in agreement.[266] The winds blew, probably with considerable force,[267] during a Dust Age over outwash fans and plains which were exposed and dried during the intervals between successive inundations. The finest dust settled in smooth sheets since its grain-size prevented accumulation in dunes.[268] Wind and water each played its part,[269] water acting locally and secondarily. Conditions were very auspicious; there were dust-storms, abundant fine silts and rock-flour, and anastomosing or braided streams which shrank or dried up in late autumn and winter.[270] The correlation of the lime content of the loess with the variations in the lime content of the local schotter, e.g. in Lower Austria, suggests that the dust from a particular source was carried to only a small distance and not over many hundreds of miles or kilometres (F. Weidenbach, 1952).

Analysis and experiment show that the characteristic grain-size (except in

loess derived from deserts, which has a poorer sorting and larger grain) arises from a twofold sifting, depending upon the rate at which grains of different size sink in water and air.[271] The sediments settled in water as flood loams and the wind winnowed the fine material. This was blown, for example, on to the northern flanks of the German Mittelgebirge and the Carpathians, and where these sank into the plains was swept far over south Russia and into the waters of the Aralo-Caspian Sea to form stratified loess, e.g. in the lower Volga.[272] Deposition gradually diminished as the winds slackened away from the ice (see p. 660). The grade is otherwise ascribed to the fact that mechanical disintegration produces dust that is eminently suitable for aeolian transport.[273]

Richthofen's researches in China (see p. 542), where glacial phenomena did not obscure the vision, led to the wide adoption of the aeolian origin under steppe conditions,[274] independently demanded by Nehring[275] and later workers[276] on zoological and phytological grounds. It was expressed for both Europe[277] and North America[278]: P. T. Virlet d'Aoust[279] had anticipated it for South America as early as 1857.

The arguments, palaeontological, topographical and structural that favour the theory are overwhelming. They include the character of the molluscs as Binney[280] early pointed out and Shimek and Nehring emphasised.[281] The shells, which are mainly terrestrial, are autochtonous and ecological[282] and delicate (even snail eggs are preserved). Some have the operculum of *Helicina orbiculata* in the aperture (the operculum becomes detached after decay sets in) and whorls are empty and devoid of clay, etc. Further paleontological evidence is furnished by the rarity of freshwater fish; by the steppe animals, e.g. *Equus caballus*, *E. ferus* (= *przewalski*) *Asinus hemionus*, *Citellus* (*Spermophilus*) *rufesens*, *C. mugosaricus*, *Marmota* (*Arctomys*) *bobac*, *M. marmota*, *Allactaga saliens*, *A. jaculus*, *Saiga tatarica*, as in Switzerland and at Thiede and Westeregeln,[283] jerboa and steppe rodents in Alsace, and the ground squirrel at Montmorency and north of the Garonne (see p. 803); by the coprolites in the Moravian loess[284]; by the bones of land vertebrates, gnawed and strewn about as carrion feeders left them[285]; and by the occasional discovery of salt-loving steppe plants.[286]

The topographical arguments refer to its indifference to the nature of the underlying rock; its great thickness near the ice-margin; its independence of altitude (loess in the Mississippi basin has an extreme vertical range of *c.* 300 m[287] and in Europe sweeps up to 500 m or more; cf. p. 517); its unilateral distribution in meridional valleys where it chiefly reposes in the lee of hills ("lee sides" are "loess sides"[288]) or on the western sides of valleys[289] (e.g. Russian Podolia, Poland, Silesia, Hungary, Thuringia, and Belgium)—while this is in general conceded, contrary inferences are drawn from the fact (see p. 532); and its absence where glacier-streams carried their muds into a sea or big inland lake, e.g. the enlarged Caspian Sea (see p. 1131) in the middle and lower Volga below Kazan.[290] The zonality is also in agreement—particles less than 0·01 mm in diameter in the Ukraine form 20% to 25%, farther south 35% to 50%, and in the extreme south over 50%.[291]

The physical characters bear equally decisive testimony. Stratification, owing to the wind's ineffectiveness as a sorting agent, is lacking (cf. p. 529); deposition was slow and each addition extremely thin[292]; the grain is fine, particularly on hill-flanks,[293] and corresponds with modern cryoconite,[294] with the size-grade fraction of the local till[295] and with dust showers and

samples collected at velocities of 18–24 miles (*c.* 29–39 km) per hour[296]; it is unweathered[297]; its molluscs are irregular in species and individuals[298]; it has rounded grains and wind-etched pebbles,[299] boulders burst by insolation,[300] a high lime-content,[301] and countless mica flakes, irregularly orientated[302]; it dovetails with wind-blown sands, as in Moravia[303] and Hungary,[304] in several German localities[305] and in North America[306]; it is buff coloured from oxidation in the air[307]; and charcoal is present in the industrial layers at Krems.[308] Lateglacial ventefacts in the Mashpee pitted plain (see p. 440) prove semi-arid and cold conditions at the time of their formation since the kettle-holes had not then formed[309]—they are not filled with sand.

Periglacial conditions are seldom found around modern ice-sheets[310] since these usually end in the sea. Yet similar dust is being raised to-day in clouds rivalling those of the desert over the *Sandur* plains[311] of Alaska, Greenland and Spitsbergen, and in connexion with the loess terrain of Asia (see p. 542), the outwash of the Rhine and other European rivers[312] and the Canterbury Plain of New Zealand.[313] Outblowing winds have sometimes, notably during dry winter seasons, etched or pitted boulders or formed dreikanter on the plains.[314] In Iceland,[315] where the dust is carried into the air to a height of 1000 m, this *Móhella* of the Icelanders, which includes volcanic ash and dust, is up to 100 m thick, encloses calcareous concretions, and sometimes wraps the whole landscape, leaving behind it a surface which is *örfoka*, i.e. unable to supply wind-borne material,[316] such as the wind-swept areas in the Antarctic, e.g. the west side of McMurdo Sound (Dry Valley), Knox Coast (Bunger oasis) and Dronning Maud Land and the "frozen lakes" (see p. 1490) of Neu-Schwabenland—this is the "pebble pavement" of Greenland.[317] The wind-blown material in Jan Mayen forms sand or snow cones.[318]

The *Steinsohle* or regular pavement of *Dreikanter*[319] (Fr. *cailloux façonnées*) ventefacts or wind-faceted pebbles (alternative names[320] are *Windkanter*, *Pyramidalgeschiebe*, *Glyptoliths* or *aeoliliths*) which forms the base of the loess in numerous European localities[321] is additional proof of wind activity. Faceted ventefacts, which have polished and greasy, fluted surfaces, sharp edges and differential etching, are also known from the glaciated area of North America,[322] where they consist of Huronian and Cambrian quartzites or of fine or medium-grained igneous rocks of the Adirondacks and New England. The pebbles or boulders, usually of hard and homogeneous rock like quartzite, granite or gneiss, may owe their ground plan to running water[323] or moving ice[324] or to the boulder's original shape,[325] as controlled by cleavage, bedding or jointing. While the shape has been attributed to the washing of till[326] or to melt-water rocking the boulders on each other (*Packungstheorie*[327]) or even sometimes as aberrant artefacts,[328] most investigators and probably all modern ones regard the boulders as a deflation residuum.[329] The wind pitted and polished the surface, as shown experimentally[330] and observed in front of modern glaciers[331] in Greenland (see below) and Iceland where winds from the ice polish cobbles, boulders and bedrock. A desert glaze has been observed about the modern glaciers of the eastern Alps.[332] Rocks of heterogeneous character give rise to ribboned or recessed ventefacts.[333]

The wind-blown dust was trapped by grasses or arctic moss,[334] as Richthofen suggested and later geologists have believed,[335] or has occasionally been observed in analogous areas to-day[336]—the northern limit of the north German loess coincided with the northern limit of such grasses[337] (see p. 533): a northern loess-tundra and a loess-steppe to the south of it has been distin-

guished, the boundary between them being a climatic and morphological one —it rises southwards into the Mittelgebirge and Alps with other climatic lines. When the blades and roots decayed from oft-repeated moistening, followed by drying and oxidation made possible by the access of fresh air,[338] they left the cylindrical hollows which traverse the loess save in aqueous varieties. The tubules bifircate downwards like rootlets[339] and are associated with living plants[340] (e.g. *Lithospermum officinale* and *Pinus*) or with black specks of vegetable matter.[341] The calcareous lining was sucked up by rootlets which still remain,[342] encrusted with such matter or completely calcified. The absence of a horizontal stratification—conifers and deciduous trees produce such a stratification by their fallen needles or leaves—and the great rarity of shells indicative of the presence of trees or bushes point in the same direction,[343] while newly made exposures in recent loess in Alaska[344] show root fibres near the surface which grade downwards into a zone where the roots are carbonised and a lower one where the fibres are gone and tube-like casts only remain. This seems to put the theory beyond doubt.

Nevertheless, some dissent because the entire profile should resemble a humus horizon[345]; the tubes have scarcely a trace of vegetation[346] (the physical nature of the loess renders this criticism nugatory), are wrong in shape,[347] extend from bottom to top,[348] and occur in marine and other deposits where the specified conditions do not prevail; the lime could not, for physiological reasons, have been drawn up by the plants but might have been deposited after their death by downward percolation[349]; the structure is either independent of the conditions of formation or is mechanical, due to climatic processes which have effaced the original differences in the various loess deposits[350]; and loess is forming to-day in Greenland and Iceland where steppe grasses do not grow. Loess, therefore, may not necessarily have been associated with steppes[351]; it may have been deposited near the treeline, i.e. between the trunks below the line and on bare land above it.[352] A uniform climatic or ecological environment is not to be expected. Thus molluscs and other fossils in the Iowan loess suggest forest in some places, grassland in others and even a progressive change from forest to grass.[353] In Illinois, molluscs in Wisconsin loess show a transition from cold types at the base to warm-temperate types at the top and an absolute increase in their number.[354]

But while it has to be admitted that the flora of the loess steppe is virtually unknown, since the loess is not suited to preserve plant remains—pine and fir with small annular rings have been found at Göbelsberg in Lower Austria, pollen and spores of various conifers and deciduous trees and especially of NAP. (steppe grasses and species of *Artemisia*) in the loess of south Russia,[355] and roots and stems in Palestinian and North American loess[356]—there must have been sparse grasses which were adapted to winter cold and summer dryness, to a short growing season in spring, and to strong winds and sudden drops in temperature.[357] The flora, however, was certainly not scarce since a considerable fauna, both molluscan and mammalian, inhabited the loess.

Deposition was probably facilitated by the dying away of the wind as the distance from the ice increased and the ice-winds were replaced by westerly winds.[358] It took place where the ground or the atmosphere had enough moisture to bind, directly or through chemical action, the particles that had settled[359] but where the area of reception was not subject to cryoturbation or solifluxion, had sufficient vegetation to prevent the dust being blown

away and was dry enough not to cause decalcification and weathering of the loess.

Although the climate may have been cold and wet[360] and deflation undoubtedly continued so long as the shifting streams kept the outwash bare of vegetation,[361], the vast majority of workers agree that it was dry though not wholly arid. Thus a dry continental climate is found in places along the margin of the ice in present-day Greenland[362] and Spitsbergen[363] (see p. 678); the loess is of the right colour and poor in alumina and humus and rich in carbonate[364] and could not have accumulated had the vegetation been thick[365]; the molluscs signify a dry climate[366] and abound to-day in the steppes of Central Asia, south-west Siberia and east Russia; and the loess loams have desiccation cracks.[367] Further (though doubtful) confirmation is provided by wind-eroded rocks,[368] chiefly sandstone, in central Europe and North America, and the shallow depressions in granite[369] that recall the wind-polished and eroded rocks, with desert glaze, observed about modern glaciers.[370]

A coating like that of deserts has been described from bedded rocks at Tübingen,[371] from Hochterrassenschotter in Switzerland,[372] and from Chellean artefacts in south France[373] and glacial pebbles in Schleswig-Holstein.[374] Arid conditions have also been postulated for Pleistocene Hungary[375] and central Europe, because of the abrupt rock-faces of the Pfälzerwald,[376] the steep, weathered surfaces of Sächsische Schweiz and adjacent regions[377] (these are partly due to the nature of the rock and to forces now operating[378]), and the ventefacts[379] of, for example, Bohemia, Brittany and south France.

The central European climate resembled that of south-east Russia[380] or north Siberia[381] to-day. Its continentality was due to the low sea-level and the continent's westerly extension[382] (see p. 1355)—aeolian action took place on the site of the present North Sea[383] and loess is found on the raised beaches of the Channel Islands[384] and the islands off Brittany down to −25 m.[385] The loess around the Black Sea has also been linked with a low level of this sea,[386] and the typical loess of northern Dalmatia, Istria and the Istrian islands with the laying dry of the northern Adriatic[387] (see p. 1225). The continentality was strengthened by ice over the epicontinental seas[388] and the close pack-ice off western Europe[389] which lowered the temperatures. It was, however, mainly related to the glacial anticyclones over the ice-sheets (see ch. XXXI) which induced dryness and much evaporation[390]—the annual precipitation in central Germany may have been *c.* 25 cm.[391] P. A. Tutkowski,[392] the chief exponent of this view, thought winds blew from the ice on to a "zone of deflation" or "zone of dreikanter", characterised by desiccated moraines, barren sands, boulder-screes, burst boulders, faceted pebbles, barchans and desert polish and coloration—the zone may have been wider in Poland than farther west, for there Aurignacian man lived farther from the ice-sheet.[393] The "glacial foehn winds" whirled the dust into a "zone of inflation" of lessened barometric gradient, the dust being trapped among the steppe vegetation. When the ice retired, the zones shifted with it: ventefacts became buried under loess and sand-loess under true loess.

The increasing distance between the younger loess and the line of the last glaciation in Germany and the Low Countries—it eventually amounts to more than 50 km—and the freedom from loess of Europe's west coast and the east coast of North America (except locally, cf. p. 519), where the loess thins

towards the Atlantic and dies out east of the Mississippi basin (over which cyclones passed north-eastwards), is readily explained by the then humidity of these countries since loess-building material was plentiful—additional factors in eastern North America were the more varied relief and the sandy nature of the drift.[394] There is no loess in Spain[395] (nor Pleistocene steppe animal in Portugal[396]), in Provence south of Lyon, or at the southern foot of the Alps in northern Italy (small patches[397] occur at Turin and near Verona) where moisture at all seasons and forests were inimical.[398] It is thin in Flanders and was formed, it is said, by *ruissellement* of heavy rains[399] in northern France where it is frequently more or less distinctly stratified and intercalated with thin beds (*cailloutis*) of sand or pebbles or angular fragments of flint.[400] In Belgium and especially in France the fashioning of sand grains and of pebbles by wind has been much less than farther north, though the action has been detected as far south as Toulouse, Bordeaux and Haute-Garonne.[401] The fact that the loess in the Rhine area is more frequently stratified than elsewhere in Germany suggests that the summer rainfall was there more plentiful[402] while the greater frequency and thickness of the solifluxion layer in England and France (pp. 1062, 1081) point to a damper and more oceanic climate and a richer supply of water in the soil[403]: damp-loving molluscan species were abundant on the high chalk Downs of southern England.[404]

In Britain, where the distribution and amount of the precipitation was not much different from that of to-day,[405] loess is either missing completely or is represented in some of the Thames brickearths,[406] especially in the Taplow Terrace north of the river, which are sometimes structureless and contain "loess dolls" and shells of *Pupa* [*Pupilla*] *muscorum*; in the loess-like loams and brickearths[407] of East Anglia, Aldershot and the Chalk plateaux at the eastern end of the Chiltern Hills—the Hallsford loess contains implements that conform closely to the Aurignacian; in the matrix of the screes[408] east of the Menai Straits, east of the Cotswold Hills, and (with terrestrial shells) about the Severn Estuary; in the red earth of palaeolithic caverns[409]; and in certain deposits in Co. Durham.[410] Wind-action in a dry climate and on pre-neolithic surfaces free from vegetation is seen in the ventefacts[411] of Godalming, Lancashire, Cheshire, Yorkshire, Lincolnshire and central England—they are common in the "head" of the Newer Drift in Yorkshire—in the desert glaze on some gravels in the Vale of York,[412] and in the sandblast surfaces at Lilleshall,[413] Shropshire. Steppe conditions are witnessed by the *Saiga tatarica*, *Ochotona* (*Lagomys*) *pusillus* and *Citellus* (*Spermophilus*) and by the wide distribution and abundance of horses in certain Pleistocene deposits, e.g. in Aurignacian caves.[414]

The age of these several features varies a great deal and cannot always be accurately determined. The sand blasting, for instance, has been referred to interglacial times[415] or the last cold phase,[416] and the faceted pebbles to a steppe climate before the submergence which brought the Vale of York drift.[417]

The widening interval between the loess and the latest drifts as we proceed westwards in Europe[418] also suggests augmented precipitation in that quarter (see above). The absence of loess from Korea and Japan on the other side of the Eurasiatic continent and from the coastal regions of western North America has the same significance, though the fine silts of Wisconsin age, associated with a foraminiferal fauna and pollen grains, revealed by cores on the outer

continental shelf of New England, have been interpreted as wind-blown loess in this region.[419] In North America, probably because the anticyclonc winds were blowing from the Atlantic and not from a continental region, as in Europe, the signs of wind action are less pronounced, and the vegetation was favoured by greater humidity[420]—the more southerly latitudes of the ice-edge may also have been a factor.

Wind direction. While westerly winds blew certainly over the Mediterranean and over the Alps, the direction of the periglacial winds has been much debated. H. Poser[421] has reconstructed the distribution of the winds in Europe during the last glaciation (fig. 105). According to him anti-

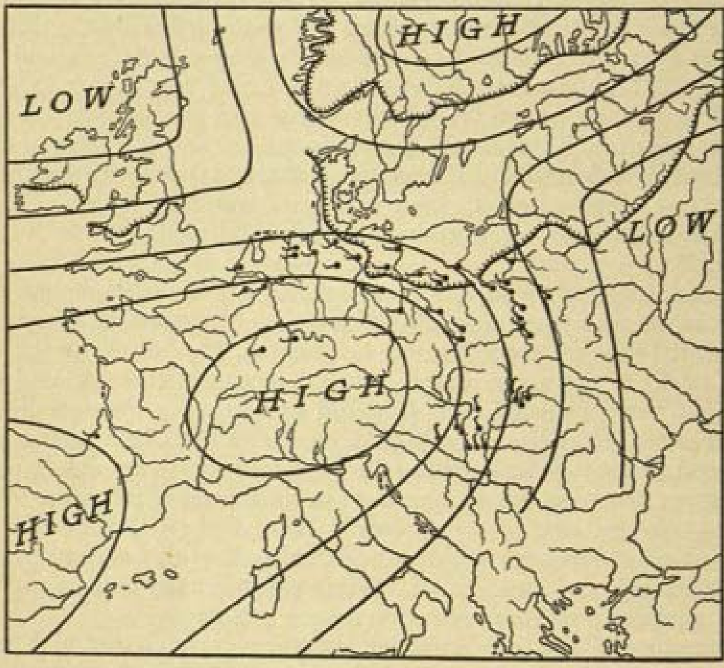


FIG. 105.—Pressure distribution and direction of winds (given by arrows) in Europe during the last glaciation. Teeth lines represent the ice-border at the beginning and the end of central Europe's lateglacial period. H. Poser, *Erdk.* 4, 1950, p. 81, fig. 1.

cylcones occurred over Spain, western Germany and Scandinavia and cyclones over the British Isles and eastern Europe. West or south-west winds blew over the Low Countries and north-west Germany, north-west winds over Poland and northerly winds over Hungary. The light winds associated with the anticyclone over eastern France explain the absence of loess from this region (cf. p. 527). Analogies with drifting snows are not compelling, though experiments[422] suggest that loess, like snow, was blown over the hills to collect in the lee. Supporters of the glacial anticyclones ascribe the loess to east or north-east "ice-winds"[423] streaming outwards from the ice.[424] Thus the loess lies on the then windward slopes of meridional valleys[425] (cf. p. 527), e.g. Oder (north of Ratibor and above Breslau), Glatzer Neisse, Elbe (near Dresden and south of Magdeburg), Elster (south of Leipzig), and Leine (near Hannover), and is absent from the wind-shadow

south-west of the Harz[426]; its source was in the east, viz. the almost desert-like interior basins of central Europe[427] (this is difficult to reconcile with the local mammalian fauna[428]), and in Russia[429] the drifts of the north-east[430] and the dry northern part of the Black Sea which was the receptacle of glacial rock-flour.[431] The loess in Russia came from the vast outwash masses of the Bug, Dnieper, Don and Volga, that on the north-east side of the German Mittelgebirge from the outwash about the *Urstromtäler*, that on the Riesengebirge from Poland, of Belgium from the dry floor of the North Sea, of Paris and Brittany from the dry bed of the English Channel,[432] of the Rhône from the Rhône Glacier, and of the Alpine foreland from the Danube[433]—the west had no source big enough. It is also emphasised that the European loess diminishes westwards in depth and area and passes eastwards into sands,[434] e.g. near Kiev and in the lower Rhine, and that the snowline was locally depressed in the Rodna Alps which received their moisture from pluvial lakes in the Dnieper.[435] The orientation of the ventefacts, e.g. in Silesia, points in the same direction.[436]

Nevertheless, it is urged that the unilateral distribution of the loess does not bear this interpretation but is related to slope without any compass orientation,[437] to the valley's asymmetry,[438] to aspect and melting snows,[439] or to the secondary action[440] of postglacial wind and rain. The asymmetry of the valleys is thus related to aspect and the unequal exposure to the sun's rays[441]—the sun thaws the ground and favours solifluxion and stream-erosion in the unfrozen ground on one side—or to the action of the wind which, by mixing loess dust with snow in the lee, also favours solifluxion.[442]

Certain facts betoken a westerly or south-westerly wind,[443] as in the Mississippi and its tributaries.[444] The sharply defined northern limit of the youngest loess in north Germany, where the loess is *c.* 1 metre thick and parallels the course of the Mittelgebirge,[445] is related to climatic factors, viz the obstructive effect of the mountains (see p. 528). Moreover, dry plains to the west supplied much of the material (see p. 523); the loess is disposed east of the rivers, e.g. the Rhine[446] (it drifted into the depression between Schwarzwald and Odenwald[447]), the Danube,[448] in Thuringia[449] and northern Bavaria,[450] and in North America[451] where this relationship holds good for the Mississippi from Minnesota to the delta, for the Ohio along its lower course, for the Illinois River from the Big Bend to its mouth, for the Missouri River as far upstream as Sioux City, and for such tributaries of the Mississippi as the Des Moines, Cedar and Iowa rivers. In any one region, e.g. Rhine valley, Russia, and the Peorian loess area of North America, there is an easterly passage from coarse to fine and from thick to thin deposits.[452] Furthermore, pumice and volcanic ash in the Rhine loess were blown farther east of the volcanic centre, e.g. in the Eifel[453] (fig. 106)—that this resulted from an upper return wind[454] is unlikely; dunes margin the eastern side of outwash plains[455]; and loess is highest on the east and lies in the lee of hills east of its source in the river valleys, e.g. the glacially swollen Mississippi. Westerly winds, in part of the nature of chinook winds,[456] are proved by the ventefacts in Wyoming,[457] by the distribution of the volcanic ash in the Great Plains[458] (see p. 979), and by the development of beaches on the eastern side of glacial Lake Missoula, Montana.[459] They blew even farther north than the Mississippi in North America[460]: dunes were heaped up by such winds in Connecticut (Wisconsin age) and south of Lake Michigan (Cary substage) and in north-west Illinois (Tazewell age). The size of the beaches in Lake Algonquin suggests northerly winds in Ontario.[461]

It has been suggested that in Europe the winds blew from the Mediterranean region and the Black Sea and Caspian Sea to the south or south-east,[462] the Mediterranean Sea because of its depth and the humidity of the air favouring precipitation.

The two seemingly opposite views have been reconciled in various ways. High mountains, e.g. the Black Forest, have the loess on their windward side while lower hills, e.g. Kaiserstuhl, collected it in their lee. Hungary furnishes other examples.[463] Alternatively, the distance from the ice was the controlling factor. Loess was deposited in north Germany by east winds, in central

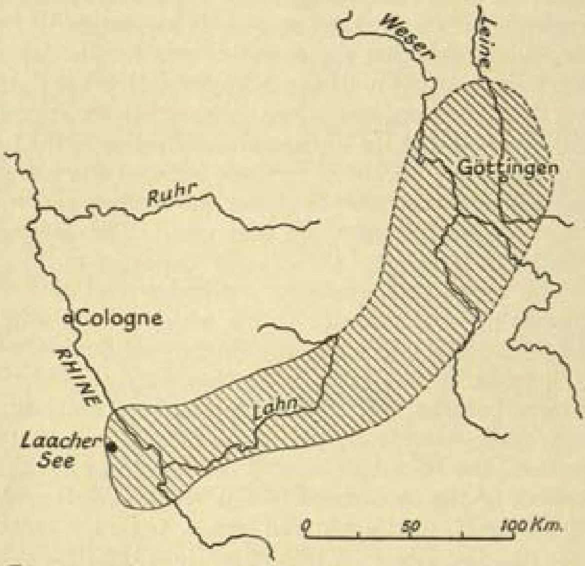


FIG. 106.—Distribution of the pumice from the last eruption of the Laacher See volcano in Alleröd time. M. Schwarzbach, *1525*, p. 66, fig. 40.

Hungary, Lower Austria, and northern Bohemia by south-east winds from the Danubian flood plain[464]; in Bessarabia, south Ukraine and in the lower Danube east of Bukarest by east winds from the exposed Black Sea floor[465]; and in Swabia, Bavaria (Franconia) and Serbia by west winds.[466] The boundary between the east and west winds has been placed as far east as the Oder and upper Elbe.[467] The loess of central Germany, like the dunes (see below), has, however, been redistributed by west winds blowing in post-glacial time.[468] In this way is explained the preservation of the loess in eastern shadow areas associated with the Mittelgebirge. In North America, as the ice-front receded northwards, the anticyclonic winds were replaced by the westerly planetary winds. Under their influence, the surface loess was lifted and carried eastwards as about the area of the Pre-Wisconsin Des Moines and Lake Michigan lobes where the loess now extends over the zone of deflation and even for a few miles on to the Wisconsin drift.[469]

On yet a fourth view, the contradictions may be removed by supposing that the loess was deposited by east winds but was redistributed by "minority" west winds which, though not prevalent, were much stronger.[470] It has also been suggested that Europe had a kind of monsoon climate[471]: east winds

blew in summer, west winds in spring and autumn when the weather was unstable, and calms prevailed in winter. Alternatively, west winds with snow blew during summer and east winds, generally speaking, during autumn and winter.[472] Only in one place in Europe, viz. east of the Rhine rift valley, did the snow and loess winds coincide. Fall winds played a subordinate part; they carried dust from the Rhône to the Côtes lyonnaises and that from the River Po on to the hills east of Turin[473] (see above).

Fossil inland dunes of Europe. Europe's fossil inland dunes have been the pivot of much of this discussion. For the most part parabola dunes, they are widely distributed, being found in north Germany,[474] notably along the *Urstromtäler* (hence thought to be river-dunes,[475]) and over *c.* 5% of the total surface (fig. 107), as between the Warthe and Netze and south of Thorn, and extensively in the Rhine—Main region from Frankfurt to beyond Darmstadt. Yet large areas are free, as north of the Baltic lake-plateau and south of the *Urstromtäler*, and even within the net as on the Lüneburger Heide, Fläming and lower Rhine. They cover wide tracts in Poland[476] where they are less bound up with the *Urstromtäler* (cf. I. Högbom's maps[477] of the dunes of Poland, Germany and Holland), and are well developed in the middle Vistula. They occur in Moldavia[478] and Russia, e.g. in the driftless territory between the Dnieper and Don.[479]

The dunes become less extensive and distinctive westwards though they are seen in Belgium in the province of Antwerp and Limburg, east of the Scheldt and on the Pliocene and Campian sands.[480] Others are known from the west coast of Schleswig[481] and from Sweden,[482] Finland,[483] including Carelia, and the Kola Peninsula.[484]

F. Solger[485] maintained that the dunes, especially in north Germany, though subsequently modified by west winds, were contemporaneous with the loess and heaped up by east winds: they are related to deflation basins and sand plains and open their crescents eastwards. But this view, which has received some encouragement,[486] is erroneous; the dunes, from Belgium to Scandinavia, Russia and the Hungarian plains,[487] were accumulated by west or south-west winds.[488] Thus the dunes as shown in plan and section are parabolic or longitudinal dunes and not barchans; their bedding is directed eastwards; and they are situated east of the sources of supply, e.g. the lower terrace of the Rhine and the sands of the Weser and Vistula. Moreover, their material becomes finer eastwards, and they wandered eastwards, e.g. in Scandinavia, as their relation to each other and to the lakes into which they advanced indicate.

The dunes are by no means all of one age, as Solger's term "dune period" would imply. They are indeed related to features of greatly differing age, such as valley sands[489] or redistributed solifluxion material[490] in the Rhine, glacial outwash in Denmark, and osar and fluvioglacial deltas of Finiglacial date in Scandinavia.[491] Yet they are unmistakably fossil; they are covered with vegetation and are at rest—except those of the lower Danube and Theiss[492]; they are deeply weathered and dark brown in colour and have pans; and the overlying beds, like the dunes themselves, contain prehistoric remains,[493] from early Mesolithic onwards. Those of the French Landes and of the Rhine, Weser and Elbe are dovetailed with terraces.[494] The prerequisite for their formation was a drier and more continental climate when the *Urstromtäler* and parts of the North Sea floor were dry[495] and vegetation was lacking or less extensive.[496] But their precise age is obscure. From

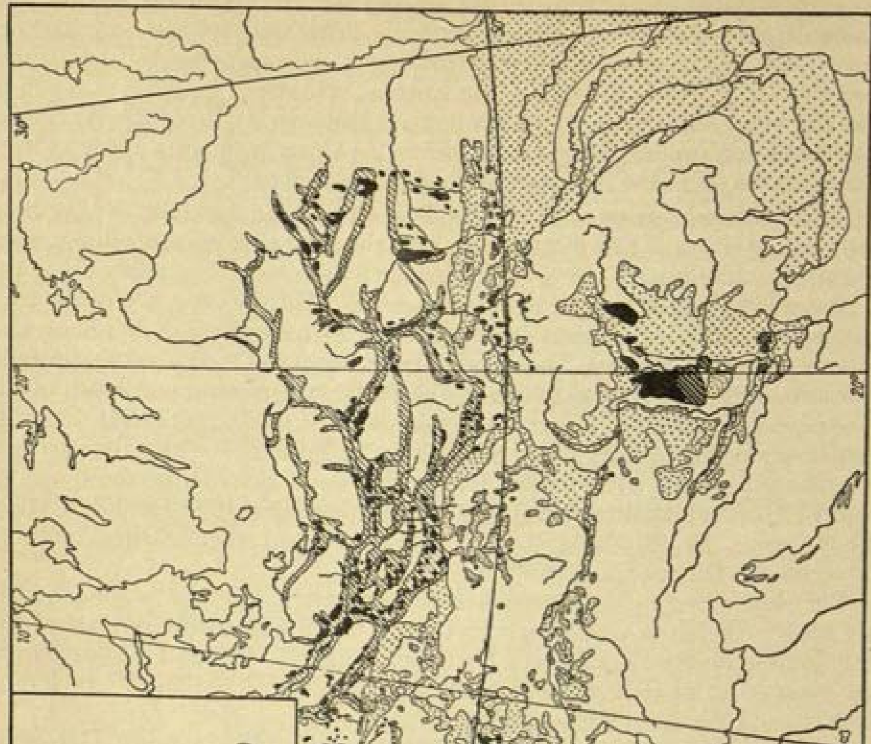


FIG. 107.—Map of the loess, fossil inland dunes and *Urstromtäler* of central and west Europe. H. Poser, *Nw.* 35, 1948, p. 270, fig. 1.

pollen analysis and their relation to the terraces of the Vistula and the Littorina shores in Finland they are placed in the Littorina period[497] or, more commonly, in the Ancylus or Boreal period,[498] e.g. the dunes in the *Urstromtäler*, when the land was higher and the water-table lower.[499] They preceded the Littorina or Finiglacial peats that overlie them.[500]

Thus the dunes, it is said, in general belong to some part of the Climatic Optimum (see p. 1482) and preceded the mixed oak forest.[501] Bronze Age finds prove that some were moving and accumulating in Sub-boreal time.[502] They were neither coeval with the loess nor periglacial.[503] Nevertheless, some of the north German dunes were lateglacial[504]; this is proved by pollen analysis[505]; by the presence of dead ice in the *Rinnenseen* at the time of their formation[506]; by the dovetailing of dunes in Holland with cryoturbation features[507] and in north-west Germany with tundra vegetation[508]; and by the lateglacial age of the loess with which they are intimately connected (see p. 539).

The inland dunes of North America,[509] which, for example, in the Mississippi valley form a close parallel with those of the north German *Urstromtäler*, were apparently heaped up postglacially by south-west winds and were probably accumulated during a drier period,[510] though they have been traced to the drying of the surface that resulted when the Mississippi and its tributaries were cut down.[511]

Interglacial age? Although the loess is forming no longer—in North America, for instance, it passes under later drifts and has normal profiles of weathering, and in Germany it is being removed so rapidly that it colours yellow the Neisse, Katzbach and Oder rivers[512]—we may ascertain its age from its geographical and stratigraphical relations to boulder-clays, outwash sheets, and terraces, or the successive mammalian faunas and palaeolithic cultures (pp. 1033, 1028). Opinion about the exact date diverges considerably. Following Nehring's work, there was general agreement[513] that the loess in north Germany and North America was laid down during the second, but more heavily and extensively during the last interglacial epoch. Thus the loess is sandwiched between moraines and drift-sheets[514] and is incorporated in the overlying drift; it includes steppe animals (see p. 527) and terrestrial molluscs of temperate species[515]; and it rests on a weathered and eroded drift.[516] Some hold that the molluscs lived in a dry climate, others that they dwelt, possibly in deep woods, in a moist and oceanic climate.[517] These disagreements suggest that the climate was not uniform and that habitat, the nature of the ground and other ecobiological conditions had an influence[518] (see p. 529).

Glacial age. The diametrically opposite view of an association with cold glacial epochs and a glacial steppe[519] or tundra[520] has found wide backing,[521] especially in more recent years and since the publication of W. Soergel's book in 1919; it marks a return to the date that was accepted before Richthofen enunciated his aeolian theory and when glacier-streams were held responsible for the deposit.

Arguments hostile to an interglacial age[522] refer to the humidity of such epochs and the lack of an interglacial mammalian fauna—*Diceros merckii* and other warm animals suggestive of an interglacial age[523] are seemingly derived[524] or, as in the case of the middle older loess of Achenheim, occur in a deposit which by its position, composition and molluscs is an interglacial slope loam.[525] Evidence of a positive nature also occurs. For instance, the last loess is periglacial to the last glaciation in the Alps,[526] north Germany and Russia,[527] North America[528] and South America.[529] Nevertheless, the occasional thin "postglacial" loess which was formed under special conditions on the last boulder-clay or moraines, e.g. in north Germany[530] (when loess

was forming up to the Pomeranian stage), Switzerland,[531] Piedmont Alps,[532] east Adriatic,[533] Scandinavia[534] and North America,[535] shows that the rule is not absolute.

This exclusion of the last loess by the ice of the last glaciation, has no meaning if the loess be not periglacial.[536] It is explained in several ways; on the aqueous hypothesis by a want of ponded waters,[537] on the aeolian theory by incoming moister conditions following the cessation of the glacial anticyclone and its foehn winds[538] (see p. 675) or the birth of the Baltic[539] or North Sea.[540] The melt-waters drained into the Baltic or in Switzerland into the *Zungenbecken* so that the glacial rock-flour settled as lake-warps and not as dry dust.[541] In central and eastern North America, the building of extensive outwash ended in the Cary substage, and the creation of the proglacial Great Lakes acted as traps for silt. The stable plains were quickly converted into grassy meadows.[542] The loess is not in contact with the terminal moraines but is separated from them by a zone—its breadth was generally 100–200 km in Silesia and west Poland 70–80 km—which was the *Sandur* and tundra zone.

Corroborative are the glacial anticyclones[543]; the similarity of the loess in its composition to the finer material of the glacial deposits[544] and its northerly passage in north Germany into glacial sands[545]; the freshness of the fragments sometimes enclosed[546]; its colour and lime-content that denote derivation from unweathered beds; its intimate association with outwash[547] and in places with glacial screes[548] and solifluxion features[549] or desert polish (see p. 1063); the subloessic drift which was eroded and weathered during the preceding interglacial epoch,[550] as is seen in connexion with the older loess of North America—convection currents, as in the modern polygonal ground (see p. 572), may have disturbed the *Steinsohle*,[551] for cryoturbation and solifluxion occurred in France and elsewhere[552] at the base of the ancient loess and of the younger loess and at the top of the ancient loess; the absence of loess nodules beneath the moraines and boulder-clays[553]; and the relationship to the ice-wedges in the ground.[554] The flora was arctic[555] and some of the molluscs, e.g. *Pupa parcedentata* and *P. columella*, had holoarctic or palaeoarctic relations[556] even as far south as Austria and Hungary. This fauna is impoverished,[557] though most of its species are climatically indifferent and of wide geographical range[558]: they indicate a "central European glacial climate"[559] rather than an arctic one (in Sweden to-day *Helix hispida* goes as far north as 62°, *Succinea oblonga* to 64°). The mammals include none of the warm southern species[560] (see above). In Europe[561] (where the loess has more abundant remains than in North America), it includes high alpine forms, e.g. *Capra ibex*, *Capella rupricapra*, *Marmota marmota*, arctic species, e.g. *Ovibus moschatus*, *Rangifer tarandus*, *Gulo borealis*, *Alopex* (*Canis*) *lagopus*, *Arvicola amphibius*, *A. arvalis*, *A. ratticeps*, *Lemmus lemmus*, *Myodes torquatus*, *M. obensis*, arcto-alpine species, e.g. *Lagopus alpinus* and *Lepus timidus*, and such cold forms as mammoth and woolly rhinoceros. In Galicia and Podolia, the July temperature was probably less than 10°C and fell northwards.[562]

The loessic molluscs in Kansas presently live at higher altitudes or latitudes,[563] and in Illinois, where most of the shells were smaller than living forms, they reveal a drier and severer climate.[564] Several species are now found far to the north, in Minnesota, Michigan and Canada, or to the west in the Rocky Mountain region. The large *Polgyra* land snails which are so

plentiful in the present fauna, especially in the warm temperate south-eastern part of the U.S.A., are rare in the loess except in southern localities. The snails, mammals (e.g. musk ox, mammoth) and trees (yew, spruce, fir, pine, larch) of the Iowan loess, together suggest a cool, subarctic climate at least four degrees of latitude farther north.[565]

The red layers in the loess of Hungary,[566] the Mississippi[567] and east Asia (see p. 542) are often attributed to moister conditions like those of the inter-glacial epochs[568] or to climatic changes attending a displacement of the poles.[569]

There are usually only two major loess horizons (see p. 1027), since loess is not associated with the earlier glaciations, except in the drier east, though an occasional badly weathered loam has been doubtfully referred to them[570]—the loess at Achenheim has been placed in the Mindel glaciation[571] and even a Villafranchian and a Donau (Danube) loess have been claimed.[572] The gumbotils of North America (see p. 901) were early identified and have recently been regarded as such ancient loess horizons.[573] Several reasons have been given for the absence of the earlier loess horizons, particularly in west and central Europe. It is explained that they were later destroyed or covered up[574]; that the glacial anticyclone did not then exist[575]; that the ice-retreat which exposed the barren ground was more rapid[576]; or that earlier drifts from which the loess could have been winnowed were lacking.[577]

Although the glacial age is now almost universally accepted, opinions differ about its relation to the climax of the cold epochs. While many espouse a correlation with the culminations[578] (full-glacial; *hochglazial*; *pleniglacial*)—each glacial period began with solifluxion and ended with loess[579]—others make the loess the concomitant of the retreat[580] or the advance[581] or place it during the maximum and the advance,[582] during the maximum and retreat,[583] before, during and after maximum glaciation,[584] or at the beginning and end of each glacial epoch.[585] These discrepancies may rest on real differences, since the loess probably continued to form until the maximum in continental Russia and central Europe[586] and later than the Wisconsin maximum in Illinois and Wisconsin and the Great Plains region of North America (see p. 972).

Loess was often the forerunner of glaciation. Thus pollen in the lignites of Chambéry shows that the moist, warm climate was followed by a colder and drier climate in which loess accumulated[587]; loess rests everywhere on weathered drifts (see p. 537) and overlies lacustrine marls in central Russia[588] and is itself overlain by the moraines of the last glaciation[589] in the Salzach region and near Lyon and Turin; and the climate during the advance, unlike that of the retreat, was probably dry (see pp. 443, 466). Yet it is objected that the anticyclone was not then established[590]; that ponded drainage prevented loess accumulation[591]; that loess, as its high lime content shows, did not spring from decalcified drift[592]; that cryoturbation forms occur only in the lower layers of the loess[593]; and that loess in Poland and elsewhere occurs on deposits of the last glaciation[594]—that there was not much such loess is because the drift had no weathered products, the climate was continental, and the periglacial zone decreased in width.

Correlation with the retreat phase, though agreeing with the fact that the various loess horizons in the Somme succeed the solifluxion nappes (see p. 1258), seemingly encounters two fatal objections; the moraines and tills of the last glaciation are generally loess-free (see above) and the melt-waters were collected into definite channels and their muds into lakes.[595]

Unfortunately, organic life was cold in all these phases and the time-value of the decomposition of the underlying drifts cannot be assessed.

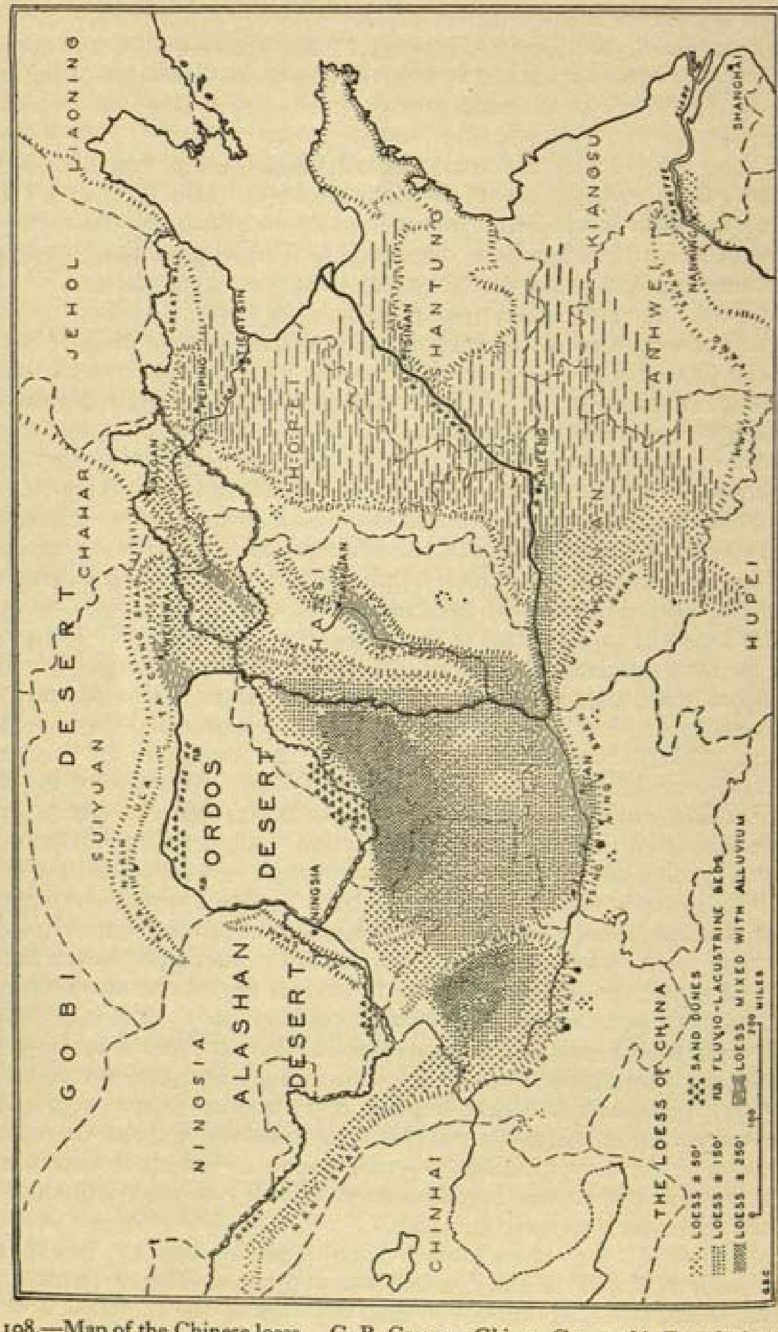


FIG. 108.—Map of the Chinese loess. G. B. Cressey, *Chinese Geographic Foundations*, 1934, p. 186.

2. *Distribution of Continental Loess*

Distribution. The continental loess is found mainly in China where it overlies rich coal-bearing beds and plays a great part in the life of the inhabitants, affecting their agriculture, water-supply, communications and dwellings. Owing to uncertain identity, its limits are as yet only approximately known, especially on the south where it becomes thin and patchy. It spreads from its north-western boundary (where it closely coincides with the line of the Great Wall) through the basin of the Yellow River and occurs at other places in Chihli, Shansi, Shensi, Honan, Kansu and Shantung[596] (fig. 108). Little has passed over the mountains of east Shansi, and Tsin-ling-shan and Tahua have provided an absolute barrier,[597] though its equivalent in central and south China is thought to be the Nanking Loessic Loam or Siashua Loam[598] or re-washed lateritic soils.[599]

The loess spreads like a blanket over the pre-existing basins. It lodges in depressions, fills up gullies, and clings like a veneer to smother the old topography of dissected plateaux and high peneplains and round the contours into smoother lines (fig. 109). Its monotonous plateaux are incised by deep river

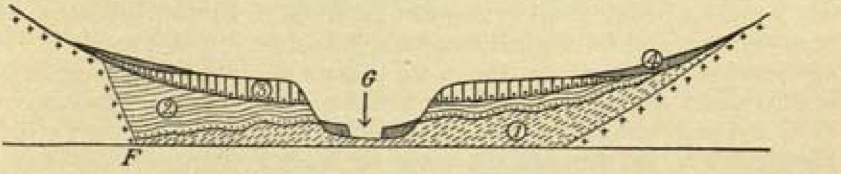


FIG. 109.—Diagram showing the structure of a Plio-Pleistocene basin in north China. 1, Pontian and middle Pliocene; 2, Villafranchian (Equus beds); 3, Red Clays; 4, Loess; G. Gorge, F. Fault. P. Teilhard de Chardin, *B. S. G. F.* 5, 8, 1938, p. 325, fig. 1 (CR. Sommaire).

canyons or are terraced naturally to conform with the subloessic unevenness or layers of loess nodules,[600] or artificially to prevent erosion and make tillage more convenient. The depth obviously varies with this buried topography and can only be gauged from exposures in youthful ravines. Estimates consequently vary considerably. Richthofen[601] gave *c.* 600 m as a maximum, a thickness which later investigators[602] reduced to 30 m, 50 m or 150 m, though recent figures[603] range up to 500 m in east Kansu, to 400–600 m in Tibet, and locally in Shansi to *c.* 450 or even 600 m with an average of *c.* 150 m. The higher figures may be exaggerations. Fruitful sources of error arise from the failure to separate loess from the overlying and more recent deposits; the absence of a clear-cut base, the loess passing insensibly downwards into older loess-like deposits; the dissected character of the subloessic surface; the re-deposited loess and alluvium which obscure low levels; and the inclusion of much material that is not true loess but beds of greater age and very different origin, lithologically, physiographically, faunistically and climatically. Nevertheless, the loess is thick in the basins and thins on to their margins.

In north-west India,[604] where the loess occurs in the Salt Range, Potwar and Baluchistan, and contains teeth and bones of large and extinct *Camelus*, *Equus*, *Bos* and *Canis*, it is at least 90 m thick.

The loess of Siberia[605] contains *Pupa*, *Helix* and *Succinea* and mammoth,

woolly rhinoceros, reindeer, saiga antelope and Irish Deer and implements resembling the Mousterian of Europe.

The absence of loess round about the deserts of Africa, America and Australia is noteworthy.[606] Small areas of fine material interpreted as loess have, it is true, been reported from Biskra in Algeria,[607] north Tripolitania,[608] Palestine and adjacent districts in Sinai[609]—in the region of Beersheba it covers *c.* 1800 sq. km, is up to 30 m thick and is still forming—the southern parts of North America,[610] Australia and South and South-West Africa.[611] The loessic nature is however disputed[612] for Algeria and Tripolitania and its absence is in general ascribed to the nearness of the sea,[613] to imperfect investigation[614] or to the lack of great flood plains which could supply the dust.[615]

Origin. The origin of the Asiatic loess has been a problem of the front rank for over seventy years. The hypothesis of water-action[616] in the form of sea or lake, suggested by R. Pumpelly[617] and C. Lyell,[618] was opposed by Richthofen.[619] In demonstrating his aeolian theory, which won general acceptance,[620] he showed that aqueous deposition was incompatible with the independence of altitude displayed by the loess (it is found up to 2400 or even 3000 m), its independence in its composition of the underlying floor, its freedom from bedding and fresh-water molluscs, its land shells, many of them very delicate, and its vertical structure indicative of plant rootlets. In his opinion, the loess was blown from the deserts of central Asia in great clouds of dust which settled on salt-steppe basins without outlet. The climate was drier and more extreme and China extended farther out into the Pacific by a barrier that lay off the present coast as far north as Korea—much of the area of the Gulf of Pe-Chihli and the Yellow Sea was then dry land.[621] Thus the loess was trapped in marginal prairies with the aid of vegetation.

Drier conditions, suggested by the remains of the ostrich (*Struthiolithus chersonensis*) and its fragile shells ("dragon's eggs"), known to the people of the latest Stone Age, are favoured by most writers,[622] including those who imagined the Hwang-ho dried up before it reached the plains.[623] G. B. Barbour[624] thought the dryness was borne out by the close conformity of the loess to the buried topography, its local passage into contemporaneous dunes, and its barrenness of life, save in the basal layers and the sandy lenses which mark ancient water-holes and stream-channels.

Many of the concretion horizons occur directly under bands of reddish-brown clay which represent old soils. The deeper loess contains a large number of these superimposed soil profiles, indicating that the loess accumulation was sporadic and not continuous.[625]

The loess, it is claimed, is still accreting in various parts of Asia,[626] in Patagonia and Argentina[627] and even in places in Europe (see p. 528)—the dust of Peking comes to-day from the Mongolian Plateau 200 miles (*c.* 320 km) away[628] (see below). Its age is said to be alluvial in Palestine[629] and neolithic[630] in Persia and in the valley of the Yenisei. However this may be (some of the dust may come from deflation of the top layers following soil erosion and intensive cultivation[631] as it does to-day in Peking[632]), its formation in general is certainly past.[633] Erosion, widespread and catastrophic,[634] is now dominant. It hollows out bridges, pipes, tubes and sinks in the loess; channels this with narrow, vertical-walled gullies, sometimes to the base, aided by seismic landslides[635]; discolours the rivers, e.g. the Hwang-ho (Yellow River), and provides the silts of this (= 40–50% by weight) and

the Yungting River of north China (90% consists of loess[636]); and causes the delta of the Yellow River to advance its front at the rapid rate of 3 km per century[637] or even one-fifth of a mile (*c.* 320 m) per annum—the Gulf of Pe-Chihli should become dry land within 2000 years.[638] This vigorous action, which has wasted the smaller hollows, trenched the deposits lower down and cut back the divides to narrow walls ("loess dykes"), results from the return of moister conditions which gave rise to "redeposited loess".[639] All stages are traceable from youth through maturity to old age, but only in the divide region of north Shansi and in parts of the Wei basin has the original smooth topography been preserved. The redeposited loess which is found mixed with alluvium and gravel on the plains is quite distinct from the primary loess of the higher country. It was laid down by both water and wind, probably in late-Pleistocene time (Panchiao stage), and entirely before the neolithic phase of Chinese civilisation.

Palaeolithic implements are embedded in the loess-cover of the Tibetan upland,[640] and Chinese neolithic artefacts above the loess[641] prove that in this country it is earlier than these cultures. Generally acknowledged to be Pleistocene[642] (Malan stage)—the Chinese loess dovetails with corrie moraines in north Shansi[643]—it was probably laid down when, owing to the greater power and dominance of the central Asiatic anticyclone of winter, the winds were stronger, drier and more persistent, and the North Pacific Low lay farther south and sucked in the winds—the hackberry endocarps associated with Peking Man (see p. 854) suggest a cool and semi-arid habitat[644] (cf. p. 1119): the summers were also cooler. The monsoon effect, owing to the low temperatures, was less far-reaching[645] (cf. p. 640), and the moisture-bearing winds from the Yellow Sea were not so deflected over north China as at present. The permanent high pressure that brought the cool dry climate of Inner Mongolia *c.* four degrees of latitude farther south, favoured rapid desiccation of the loosely consolidated and finely divided soil when the binding vegetation was removed.

To link the Pleistocene events and physiographic changes in China with Europe's glacial succession is necessarily difficult: the two regions are separated by the width of a continent. Animal remains and fossiliferous horizons are few—the Siberian loess contains remains of *Rangifer tarandus*, *Cervus megaceros*, *Saiga tatarica*, *Bison priscus* and *Equus caballus*[646]—and the workmanship of the human implements presents too many differences. We are ignorant of the exact relationships of the faunal groups in the two regions and still more of the palaeolithic cultures. Nevertheless, the almost unbroken continuity of the Chinese loess with that of Europe,[647] and the occurrence in both of the woolly rhinoceros suggest that the loess in the two regions was contemporaneous. In China it was almost certainly upper palaeolithic,[648] typology suggesting Mousterian and Aurignacian affinities,[649] and a crude upper palaeolithic industry occurs in the loess of the upper Yenisei.[650] Barbour[651] distinguished two loess periods coincident respectively with the Mindel glaciation (= later Sanmenian) and the Riss-Würm twin glaciation (= Malan stage), though others place the entire loessic event in this twin glaciation,[652] refer the Malan loess to the Würm alone,[653] or differentiate between a Malan or Riss glaciation and a Würm glaciation represented by the Mongolian Sands[654]—the Sanmenian is then correlated with the first, and Chou-K'ou-tien with the second glaciation. Peking man, however, has been placed in an interglacial epoch.[655]

Although the basal conglomerate and stratified beds with their *Limnaea*, *Planorbis*, *Unio* and other fresh-water molluscs are definitely aquatic, the Chinese loess is mostly aeolian. It was laid down when the climate was drier (see above) and colder,[656] probably during the winter,[657] and when the land was higher and more extensive[658]—the loess descends to 35 m B.S.L.[659] The shore-line, it has been said, was displaced *c.* 1100 km off north China.[660] The plants in the Lower Sanmenian indicate a cool, semi-arid climate.[661] The mollusca,[662] which are usually fresh-looking and perfectly preserved, are generally Palaeoarctic in character, and essentially similar to those now living in north China. The genus *Cathaica* was dominant then as now and *Bradybaena* was also important.

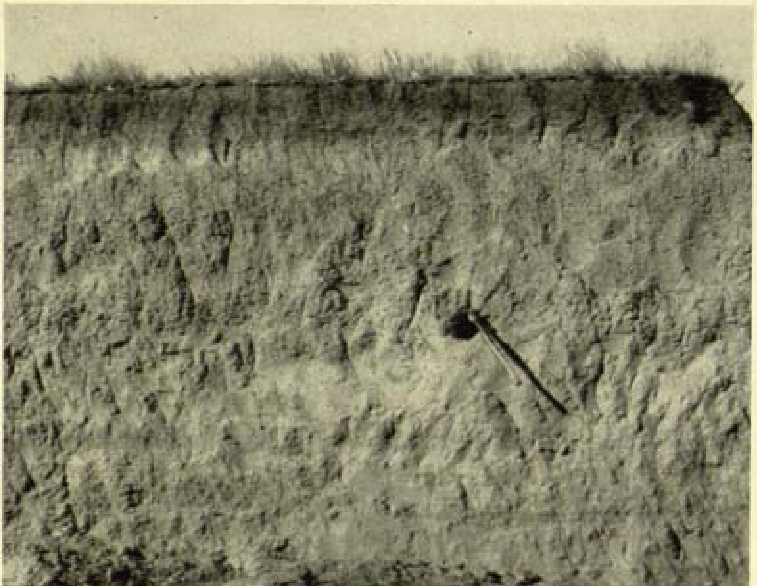
The importance of water in building up the Asiatic loess has been repeatedly stressed in recent years.[663] J. G. Andersson[664] thought the loess was fluviatile since it was thickest along the ancient valleys and was interbedded with gravel. Others,[665] including B. Willis for China[666] and W. M. Davis and E. Huntington for Turkestan,[667] have emphasised the interaction of wind and water. Aggrading rivers, assigned an almost continuous though variable activity, spread their loads over the confluent flood-plains (the sites of loess basins), the silt being sorted and re-sorted in alternate seasons.

The source of this prodigious quantity of material which spans the whole width of Asia is perplexing. While Richthofen stressed the local origin of much of the Chinese loess, and the marked concentrations near the Ordos and Tarim basins suggest that these rather than the distant Gobi were the source,[668] we must look much further,[669] as Pumpelly early recognised,[670] and seek the home of the loess in the areas of wind-erosion in the central Asian hinterland. The immediate breeding place was either the muds of Pleistocene lakes, e.g. Lob Nor, the flood-plains of rivers[671] including the glacially swollen Hwang-ho and its tributaries,[672] or the glacial outwash from Pleistocene glaciers in central Asia[673]—four loess horizons, corresponding to four glaciations, have been postulated for India and Burma.[674] It may also have been the basin sediments and vast Tertiary accumulations of secular decay,[675] as in Takla-Maken of Sinkiang, or the readily disintegrated rocks belonging to the Mesozoic and Cainozoic,[676] and the material, including talus, covering the slopes of hills and mountains, silty flats, the shores of lakes and the banks of rivers. In central Mongolia, the floor often consists of poorly consolidated Mesozoic and early Tertiary sandy-basin deposits, and an enormous mass of material has certainly been removed from the country in late geological time.[677] Climatic change towards greater aridity killed off the surface vegetation and released vast quantities of finely disintegrated debris to which smaller amounts of glacial, lacustrine and fluviatile detritus were added locally.

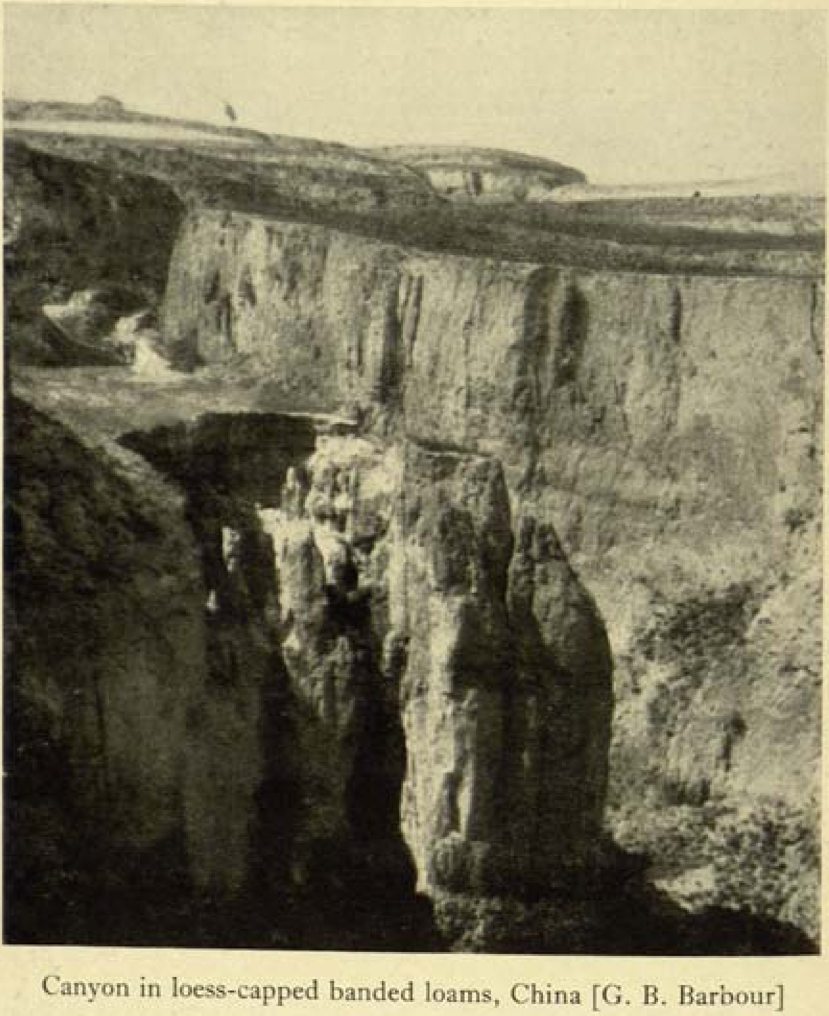
The lightest material was carried farthest (= Chinese loess facies) and the heavier was dropped earlier as sand dunes in the more central regions, e.g. in Mongolia (= Mongolian facies[678]). The lime content in Shensi is generally highest in the relatively arid regions near its supposed northern source, on account of the more rapid decomposition of the loess, the smaller admixture of siliceous material, and the limited rainfall and leaching of to-day.[679] The loess migrated by stages and at first probably did not cover all its present region.

What was formerly thought to be a single unit is now known to be a complex of superficial deposits representing four or more epochs of accumulation separated by intervals of erosion. The basal red beds are widespread in

A. Loess with two *Leimenzonen*, Paudorf, south of Krems, Lower Austria [G. Götzingen, transmitted by J. Fink]



B. Loess of Wisconsin age with two soil zones, Buffalo County, South Dakota [R. F. Flint : U.S. Geol. Surv.]



Canyon in loess-capped banded loams, China [G. B. Barbour]

north China and south-east Mongolia.[680] As in South America, they are distinguished from the loess by their colour, their slightly more clayey texture, and their less perfect vertical structure. In the centre of the basins they grade into stratified beds or laminated lake-silts with fresh-water molluscs, plant fragments and occasional gypsum layers.

In China, these beds were deposited on Pliocene (Pontian) steppes,[681] as is suggested on physiographic grounds and by the Hipparion fauna (*Hipparion*, *Sinotherium*, *Machairodus*, *Stegodon* and other genera[682]). They succeeded a late-Pliocene "lateritic period"[683] which characterised all central and southern China but encroached on the Pleistocene in the south.

Later in age, representing possibly the warm Pliocene-Pleistocene transition of Europe,[684] are the brickearths of the Laoho phase,[685] and the sands and gravels of the Sanmenian of Shensi and north Chihli whose lowest layers, the Quadrula Sands (discovered by K. V. King[686] in 1918 in the San Men Rapids on the border of Shansi and Honan provinces), were partially deposited in waters ponded by tectonic movements.[687] Altogether, they were roughly coeval with the lacustrine facies of the Nihowan substage,[688] the torrential facies of Sangkan-ho and the red loessic facies.[689] The loessic loams and boulder-bearing conglomerate at the base of the Lower Sanmenian in north China and the plant beds of the Taiku basin, which yield a cool, semi-arid flora, have been correlated with the first (Günz ?) glaciation.[690]

The Sanmenian, the oriental facies of a large central Eurasiatic faunistic association, has warm mammalia of Villafranchian age, including *Hyaena* cf. *sinensis*, *Elephas* cf. *namadicus*, *Diceros* cf. *merckii*, *Machairodus*, *Elasmotherium* and *Meles* among other genera,[691] and some large fresh-water bivalves.[692] The Villafranchian north of Tsinling[693] shows a marked increase of modern types, including forms from south Asia, e.g. bison, or from North America, e.g. *Equus sanmeniensis* and *Paracamelus*, or from the west, e.g. *Cervus boulai*, or evolving locally, e.g. *Canis*, *Ursus*, *Lutra*, *Lynx*, *Elasmotherium*, *Diceros merckii*, *Mammuthus planifrons*, *M. meridionalis* and *Elephas namadicus*. There was a massive disappearance of Pliocene forms including giraffes, *Stegodon* and *Mastodon*.

The basal beds of the overlying Chou-K'ou-tien series of humid and lacustrine origin are separated from the Nihowan stage by the Huangshui erosion phase which marks the Plio-Pleistocene boundary in north China (see p. 600), though nearly all the archaic types so characteristic of the lower and middle Pliocene, such as *Hyaennactos*, *Ictitherium*, *Chilotherium*, Giraffidae, Mastodontidae, etc., were extinct before the dawn of the Nihowan age.[694] They contain, besides a remanié fauna of Sanmenian age, typical Pleistocene mammals[695] of an Asiatic temperate steppe character, e.g. *Tichorhinus antiquitatis* (?), *Elephas namadicus*, *Equus hemionus*, *E.* cf. *namadicus*, *E.* cf. *przewalskii*, *Camelus knoblochi*, *Crocuta spelaea*, *Ursus*, *Meles taxus*, *Cervus elephas*, *C. megaceros* var. *mongoliae*, *Struthio*[696] (and its fossil egg shells), gazelles and antelopes—*Struthio* ranged in China from Pontian into the loess period. The fossiliferous localities include the breccias of Tshéli, notably that of Chou-K'ou-tien, 42 km south-west of Peking, which yielded deer and other animals,[697] fish, amphibia and reptiles,[698] remains of Peking man[699] (*Sinanthropus*) and a mixture of Mousterian, Aurignacian and microlithic industries, incised bones, and horn and bone harpoons like those of the European upper Palaeolithic.[700] Differences in the raw materials—at Chou-K'ou-tien these consist of vein quartz, schist, quartzite, flint, volcanic rocks and

limestone—make a detailed correlation between the industries of China and Europe at present impracticable.

This Choukoutien phase, which seemingly demands a moister and warmer climate, is linked physiographically more closely with the overlying loess but faunistically with the underlying Sanmenian. It has been referred to the Günz-Mindel interglacial,[701] to the closing phase of the Mindel glaciation[702] or to the Mindel-Riss interglacial.[703] The Malan loess has an upper Pleistocene mammalian fauna[704] of *Crocuta crocuta*, *Tichorhinus antiquitatis*, *Euryceros ordosianus*, *Spiroceras*, *Elephas namadicus*, water-buffalo, and camel. *Hemionus*, *Elephas*, true *Bos*, and *Mammuthus primigenius* (in Manchuria) appear for the first time, and *Hyaena sinensis*, *Machairodus*, *Diceros merckii*, *Equus sanmeniensis*, *Trogontherium*, *Siphneus tingi* and *S. arvicolinus* have disappeared. The Ordos industry, with its points, scrapers and borers of quartzite, contemporaneous with *Elephas namadicus*, *Tichorhinus antiquitatis*, *Bos primigenius*, etc., belongs to the upper part of the loess.[705]

In the last stage of this major cycle, post-Panchiao, almost all the Pleistocene forms disappear and only recent forms, nowadays living in other areas, are found (Upper Cave, Chou-K'ou-tien). The Panchiao erosion denotes a marked climatic change which may correspond with the last glaciation in Europe.[706]

The blackearths of north Jehol, south-east Ordos and Manchuria cover the period from the upper Palaeolithic to Neolithic which saw the replacement of the Pleistocene fauna by the modern fauna.[707]

The Sinanthropus beds are linked by a chain of *Stegodon*-bearing fissures extending from the Yangtse-kiang basin through Indo-China with the Trinil beds of Java.[708] They were probably contemporaneous with the Boulder Conglomerate, of Upper Siwalik age and glacial outwash origin, of northern India (which also contains the earliest human records, viz. artefacts of quartzite and other metamorphic rocks[709]). China, Java and India at this time all had Siwalik animals and were characterised by the appearance of *Elephas namadicus* (see p. 545).

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Chapter XXVII

CRYOTURBATION

The processes of weathering in polar regions differ strikingly from those in lower latitudes. Mechanical transport by running water is, for example, greatly minimised by the scanty rainfall, by frost and at lower elevations by plant growth. Chemical action is very subordinate; inimical are the snow, the frozen ground, and the cold summer months when the sun's heat is mainly used up in melting the snow and thawing the ground, and when water circulates only to a very shallow depth. Nevertheless, chemical action is not quite unknown in the drier lands bordering the glaciers. In the steppe-like areas of west Greenland,[1] there are salts, salt effervescence and salt lakes without outlet. Salt effervescence has also been noticed in Spitsbergen[2] and in Antarctica[3] (South Victoria Land and Gaussberg). Soil analyses[4] show that, partly as a result of organic substances, chemical weathering including the oxidation of the iron and the removal of the carbonates occurs in east Greenland and in Spitsbergen where it is aided by the frozen ground which prevents percolation and so causes a concentration of organic acids and of mineral salts in the suprapermafrost water. Yet the sands in polar regions correspond in composition and colour to the mother rock; felspars, as on Mount Erebus, are remarkably fresh and gossans are thin and rare.[5]

Some or most of the chemical weathering in the Arctic may date back to the postglacial climatic optimum[6] (see p. 1482). Higher temperatures also enhanced it in the periglacial zone of the Glacial period[7]; for experiments show that chemical action must have taken place.[8] Yet in general the schotter in central Europe differed little chemically from the Pliocene material; the less resistant minerals and fragments are still preserved.[9]

Insolation is also important in polar weathering.[10] But the chief agency in the periglacial zone is unquestionably frost; its geological effects were and are very widespread and considerable. The term cryoturbation, first suggested in Holland[11] for these important processes, has now been widely adopted.[12]

1. *Permafrost*

In territories where the intense cold freezes the ground-water and arrests the flow of springs, the ground is permanently frozen. This frozen soil (Ger. *Eisboden*, *Frostboden*, *Dauerfrostboden*; Swed. *tjäele*; Russ. *Merslotá*), the *permafrost* of S. W. Muller and *pergelisol* of K. Bryan,[13] has been recognised in Siberia since the end of the 17th century. It became of wide interest towards the end of the last century during the construction of the Trans-Siberian Railway and the gold rush to the Yukon, and later during the excursion to Spitsbergen of the 11th International Geological Congress in 1910.

Frozen ground is not restricted to loose soil: bedrock is also frozen. The lower limit, which is probably uneven owing to ground water circulation and

the varying conductivity of the rock, is R. Pohle's *Niefrostboden*[14] which is either dry or holds water under artesian pressure[15]—for this (and other unfrozen ground) W. S. Muller has suggested the Russian word *talik* and K. Bryan the word *tabetisol*. Its higher limit is the surface except during summer when the top layer thaws. This thaw layer (Ger. *Auftauboden*; Swed. *tjällossning*), the *Mollisol* of Bryan and "active layer" of Muller, is of importance agriculturally since it supplies the crops with sufficient moisture. It has a depth which A. v. Middendorff estimated at 25 m but is more probably in the northern tundra 1·5 m, in the Siberian climate of L. S. Berg up to 5–7 m, in Spitsbergen[16] 0·6 m, in north Sweden[17] at least 2 m and in Manitoba[18] 2·5 m. It varies with drainage, altitude and aspect, the snowfall, the texture of the soil and the character of the vegetation—it is at a minimum in peat or highly organic sediments and increases successively in clay, silt and mud, to a maximum in gravelly ground or exposed bedrock. The lower surface, the frost-table of Muller, is uneven.

The depth of the permafrost, which may be ascertained by electromagnetic or seismic methods,[19] varies between 0·43 and 69·78 m and averages 6·55 m[20] and is related to the annual air temperature as follows[21]: 1–2°C, 20 m; 0–1°C, 50 m; below 0°C, more than 80 m. Greater depths are, however, known (see below).

During winter, the lowering of the temperature in moist ground proceeds downwards at an irregular rate. Much moisture retards the penetration of cold because of the latent heat of fusion, so that the ground remains for a considerable time above freezing point. This is the *zero-curtain* of Sumgin. Near the base of the active layer, the zero-curtain may last more than one month or for a few months. The depth of the zero-curtain depends among other factors upon the saturation and heat conducting of the ground and the insulating effect of vegetation and of snow.

In autumn and winter, after the soil has frozen, water is trapped between the "wet" frozen ground below and the frozen crust above. Its high pressure raises the tundra as a gigantic blister,[22] up to 6 m high, from which water spurts out in geyser fashion. The round mounds[23] or "hydrolaccolites", the *pols* of north Sweden, the *Náledj* of Siberia, the *palsen* (sing. *pals*) of northern Europe, and the *pingos* of arctic America, whose height averages 12–35 m but may be 90 m, are split open by an irregular system of fractures and have a well-defined stratification dipping outwards. They may have been produced by erosion of a plain or of drifting sand, but more probably by hydraulic pressure or by local upheaval due to expansion following the progressive freezing of a body or lens of water or of semifluid mud or silt (enclosed between bedrock and frozen surface soil) and the force of crystallisation. Later they decay and collapse, summer melting being vigorous once the cracks have been formed, and leave behind surface hollows, pits, sinks or small lakes, to form the *Dauerfrostbodenkarst*, *Thermokarst* or *Gefrierkarst*[24]

Distribution and origin. The frozen ground, Pohle's *Gefrornis*, P. J. Koloszkow's *kryosphäre* and the permafrost of recent American literature,[25] occurs on an unknown scale in the Antarctic and is of vast extent in the colder lands of the northern hemisphere, namely, in polar, continental and alpine or high mountainous areas: the various designations given to it, namely, periglacial (see p. 510), subglacial,[26] arctic,[27] polar[28] or subnival,[29] have reference to these distributions. About one-fifth (22%) of the land area of the world

is underlain by it.[30] It occurs in Alaska,[31] where in shafts it is up to 60, 152 or even *c.* 300 m thick, underlies most of Canada,[32] including the tracts east and west of Hudson Bay (but not the bay itself[33]), and is occasionally found in more southerly parts of North America, e.g. in Connecticut.[34] The North American boundary follows the course of the Yukon, then turns eastwards along the 60th Parallel to the 110th meridian where it swings south-eastwards to the southern end of Hudson Bay and then to the north-east (fig. 110): it is said to coincide with the annual mean isotherm of −5°C. It floors the *Yderland* of Greenland (though not in the south and south-east[35]) and occurs in Spitsbergen[36] where it is 230 m or even 320 m thick in the coal mines, but is absent from big glaciers and lakes and from the sea-bottom[37] to 100 m from the coast. In Bear Island[38] it may be 60–70 m deep and in Novaya Zemlya,[39] as proved by seismic methods, *c.* 300 m. It is absent from Iceland.[40]

Most of Europe is outside its bounds. Where it has been discovered, e.g. in north Scandinavia,[41] as about Dovrefjeld and Torne Träsk, it is very thin; in Finland it was 1 m deep.[42] Seasonally frozen ground or *pereletok* occurs outside its limit, and exceptional winters create a temporary frozen ground in other localities.[43]

The greatest extent is in Siberia where J. G. Gmelin[44] in 1752 found the ground was frozen 100 ft (30 m) down. For long the maximum depth encountered was 116·5 m in the Schergin shaft, Jakutsk[45] (since deepened to 145 m), though the temperature of −3°C at the bottom suggested an unpenetrated depth which, it was calculated, would bring the total in this locality to 200 m.[46] A new maximum of over 274 m has since been found at Anderma in north Siberia[47] (69° 50′ N. Lat.; 62° E. Long.), and a figure of 400 m has been mentioned for Nordwik in the Jakutsk area[48] and for the New Siberian Islands,[49] and of 800–1000 m between Cape Chelyuskin and the mouth of the Indigirka.[50] Unusually thick permafrost tends to occur in areas which were not glaciated during the Pleistocene.[51] Frozen ground and Pleistocene glaciation have often been said to be complementary (see p. 1456), though this may be doubtful.[52] The depth, as is shown by the shape of the temperature curves, may be increasing (aggradation or pergelation) in some places and decreasing (degradation or depergelation) in others[53] (see p. 148).

The distribution of the permafrost in Asia is still very imperfectly known: much information regarding it is given in the *Arb. Kom. Dauerfrostboden Ak. Wiss. U.S.S.R.* 1931→. Schostakowitsch[54] thought it underlay 6 million sq. km between 100° and 135° E. Long. and north of 50° N. Lat. E. W. Malchenk[55] described its distribution in east Siberia and M. J. Sumgrin,[56] from a comprehensive survey, estimated its area in the U.S.S.R. at 9,658,000 sq. km or 45% of the state—A. A. Grigor'ev[57] gives a figure of 48%.

Each investigation since Middendorf's (1862) has thrust the boundary farther south. Its course is still obscure and is drawn in widely different ways. This is because the continuous Gefrornis embraces island-like patches of unfrozen ground or *talik* of unknown extent, has numerous sporadic outliers of equal indefiniteness, and may be covered with wood and forest.[58] In Sumgrin's reconstruction, the boundary runs irregularly from the White Sea along the 65th Parallel and proceeds into Mongolia, passing north-eastwards to meet the Bering Sea in 58° N. Lat. It extends to the Arctic coast and beneath the sea, e.g. Khatanga, but generally thins abruptly to the north under the Arctic Ocean. The *Great Soviet-World Atlas*, Moscow, I,

1937, also gives a map of the distribution.[59] The thickness in general diminishes from north to south.

The seasonal changes of temperature in the permafrost decrease downwards to a horizon of constant temperature below which the temperature begins to

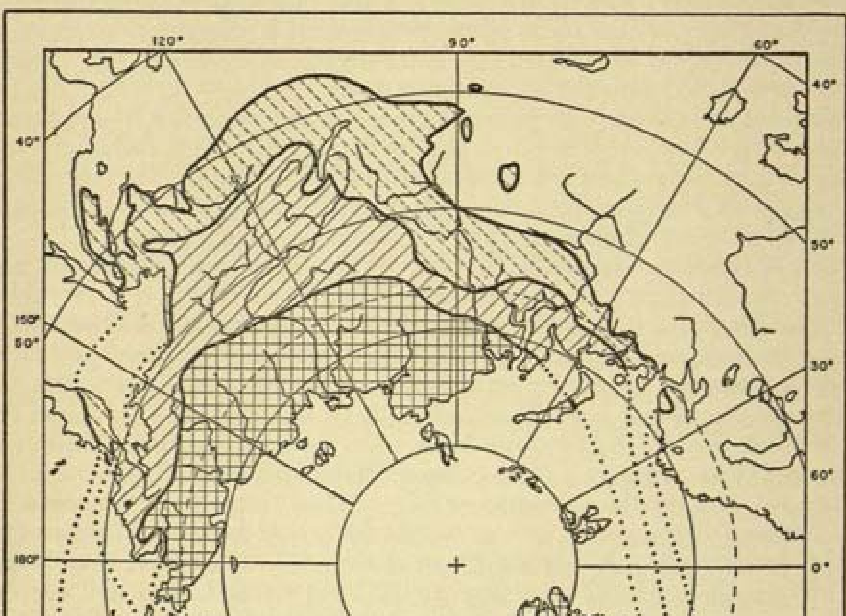

FIG. 110.—Distribution of the permafrost in the northern hemisphere. Cross-hatching, continuous permafrost; oblique shading, discontinuous permafrost; broken lines, zone with islands of permafrost. R. F. Black, *1868*, p. 275, fig. 1.

rise. Where the temperature sinks below 0°C in winter only, the frozen ground is limited to this season and is shallow. If the temperature is below 0°C all the year round the ground remains permanently and deeply frozen. But the matter is by no means so simple as this. Ground drainage, altitude —it is somewhat more in hilly ground[60]—aspect, vegetation, e.g. forest or

moss, texture and composition of the soil, seasonal air temperatures and winter snow all exercise an influence.[61] The beds of swamps near Fort Nelson are unfrozen although that place has a mean annual temperature of −6·2°C. The ground beneath the Mackenzie and Churchill rivers is also unfrozen. The geographical limit in lowlands has been drawn at the annual isotherm of −2°C[62] or −4° to −6°C,[63] alternatively at the 25-mm precipitation line of the three winter months.[64] The former neglects precipitation and snow which are very strong forces in Siberia.[65]

Asia, according to Schostakowitch,[66] has two great areas of frozen ground, a small one in the north and a larger one enclosing the greatest depth, in Transbaikalia and west of the Amur (there are outliers[67] in east Tibet, Muztag-Ata in Turkestan, and in the Pamirs and Altai). Its distribution in his opinion depends upon the mean winter temperature and the depth of the snow in January or the quotient of winter temperature divided by snow thickness (measured in centimetres). Absent if the quotient is less than 0·5, it may thus be missing if the snow is thick and the mean annual temperature is −7·4°C, but it may form if the snow is less than 10 cm deep and the temperature only −1·7°C. Sumgrin[68] criticised these conclusions because the permafrost is thickest in north Siberia, the coefficient ignores other vital factors, and no criterion exists by which to predict the presence or thickness of the frozen ground.

The latter may be a relic of the Glacial period.[69] Thus it is very thick; its upper surface is in places deep (30 m) within the ground[70] and far below the level of freezing to-day; and it encloses large masses of buried ice (see below) and frozen mammoth and other animals (see p. 648) and sometimes imprisons an unfrozen water-bearing layer.[71] Moreover, numerous *Einsturzseen* and funnel-shaped hollows, including the orientated lakes, up to 6 m or 18 m deep, abound in northern Alaska[72] and in coastal north-east Siberia[73]; at the Skovorodino station of the Transbaikal railway the temperature fell downwards[74]; and the downward temperature gradient does not correspond to the present climatic conditions.[75] The prevalent anticyclonic conditions favour not glaciers but frozen ground.

Whether or not the frozen ground is now stationary or diminishing is undecided. The upper limit has considerably receded since Middendorf's observations a century ago and the suction of the atmospheric air into boreholes, sometimes noticed, may point to the melting of intermittent ice-layers or ice-cement and the creation of voids.[76] The frozen ground is decreasing from the base upwards in parts of North America and has disappeared recently in places in Greenland.[77] On the other hand, expansion or aggradation of the frozen ground is taking place in the growing deltas of the Arctic Ocean,[78] including that of the Mackenzie River, and in recently built river islands and bars in north Siberia.

Nevertheless, there is an increasing tendency to think with G. Wild (1882) and Schostakowitch[79] that the permafrost corresponds to present conditions[80] and, though shrinking in many places, e.g. in the Kola Peninsula, in parts of Alaska, in the Petschora region and in other parts of Siberia (see p. 148), is stable and not gradually shrinking towards final disappearance as the rival theory would require. Thus the permafrost occurs in railway embankments; plants are not preserved in the older terraces of the Amur basin[81]; neolithic remains are found in the frozen ground of Mongolia[82] and cones and wood of Siberian larch in that of the Obi region (see p. 470); and

the permafrost of Greenland encloses Eskimo costumes and coffins, pierced by and matted with roots.[83] Furthermore, while the ice-sheets have melted the frozen ground has not thawed; the areas of permafrost and of Pleistocene glaciation are not complementary (see above); snows probably protected much of Pleistocene Siberia[84]; and extensive glaciation, e.g. in North America, seems to relegate the frozen ground to postglacial time (cf. p. 1456). The unfrozen *talik* on this theory is the lower part of a former active layer formed during a warmer period and left unfrozen when a later or the present cold climate caused the active layer to refreeze almost to its base.[85] The present limit therefore is probably the line to which the Pleistocene permafrost has shrunk or expanded under present conditions. A distinction has been made between permafrost which contains mammoth carcases and is a relic, and ground which does not and corresponds to present conditions[86]: permafrost which reappears when destroyed has been called active, that which does not reappear, passive.

Buildings and earth-fills alter the local thermal régime and cause the frost-table to change its position.[87] In the so-called passive method of working, peat and other surface material are left intact and additional insulation is provided so that the thawing of the ground does not take place. The active method is used where the permafrost is thin. The ground is either thawed out or that which is susceptible to swelling is removed. Areas with excessive water are excavated and the seepages are drained. Firm foundations may be obtained by anchoring with piles driven by steam points into the permafrost. Structures resting on clean gravel or sand of 20 ft (*c.* 6 m) or more in thickness are not damaged by settling or heaving since ice-lenses do not grow in them. The main problems arise if there is a thick layer of fine-grained soil or if layers of sand and silt alternate. Permafrost has influenced the customs and uses of the inhabitants, e.g. by hindering the burial of corpses which may therefore be placed under stone heaps or in boxes, and to-day is most important in constructing a modern city with foundations, roads, railways, bridges, airfields, water-mains, water-supplies and other services.[88] Pipes in the *talik* are unlikely to freeze or break, though some *taliks* flow like molasses.

Permafrost affects the vegetation by its influence on drainage and by repressing all deep-rooted species and limiting growth to those which have shallow roots.[89] Trees, like pine, with permanent tap-roots, do not thrive on permafrost. Larch, tamarack, black spruce and birch, however, can exist since they either have shallow roots or form auxiliary roots. Dwarfed or stunted birch usually indicates a permafrost close to the surface. Were the soil however not so waterlogged it would probably revert in both North America and Siberia to barren desert on account of the climate.

2. *Ground-ice*

Character and distribution. The permafrost in some regions encloses ground-ice (Ger. *Bodeneis*; Swed. *jordbundsis*; Russ. *Kamenny*), especially on the cooler sides of the valleys.[90] This ice, which occurs as thin films, grains, veinlets, large vertical wedges or horizontal sheets, and irregular masses of all sizes, varies in colour from pure white to blue, green, brown or yellow. Its structure[91] too varies with the environment and undergoes changes under pressure: it may be prismatic or granular, contain numerous air bubbles which are either orientated or without orientation, and have horizontal or

vertical planes. The solid, motionless masses, penetrated by cracks filled with loam and sand, are sometimes very deep; 300 ft (*c.* 90 m) was found in Kotzbue Sound, Alaska, and more than 320 or 365 feet (98 or 111 m) elsewhere.[92] The greatest depths occur almost exclusively in the New Siberian Islands, at the mouths of the big rivers on the adjacent mainland, and in Eschscholtz Bay, Alaska.

Ground-ice was first discovered in 1816 by O. v. Kotzebue[93] on the north coast of Alaska and later in north Siberia by Middendorf[94] on the Taimyr Peninsula and by E. v. Toll[95] who named it *Steineis*. Since then it has been found widely distributed[96] (see fig. 101) within the horizontal and vertical limits of the frozen ground and where the mean annual temperature is 3·9°C or less.[97] In Alaska,[98] it margins the whole eastern shore of Bering Sea and Strait and is widespread a few feet below the tundra. In Siberia,[99] it is found in valleys, e.g. the Jana, Indigirka, Kolyma, Lena, Chatanga and Ob, in imposing coastal cliffs, and under the straits between the off-lying islands. Here, overlain by clay and turf and underlain by till, it is separated into two horizons by a layer containing plants and animals (see p. 649), though some deny this double occurrence or deem it of no particular significance.[100]

Ground-ice has also been described from Herschel Island[101] off the north Canadian coast, from Novaya Zemlya,[102] north Fennoscandia,[103] Mongolian Altai,[104] Pamirs[105] and from Spitsbergen[106] (in beaches at Coles Bay it is up to 15 m thick and 4 km long) where its bedded layers underlie the flat floors of marginal terraces and valleys.

Date and origin. The date and origin of the ice are but imperfectly known; for though each hypothesis may explain a particular case, none has won universal application. In all probability, the ice arose in several ways,[107] e.g. by sublimation and condensation,[108] and by the freezing of waters that have descended,[109] in veins and along the bedding, to beneath peats and frozen ground. This was notably so in flat terrain where the mean annual temperature is low, viz. 4–6°C, the ground drainage is almost or quite stagnant, and the rocks have low conductivity and considerable absorbing powers. It may also arise from the freezing of lakes,[110] of lagoons and bays,[111] or of waters ascending veins and bedding planes (Pohle's ice-lenticles and J. B. Tyrrell's crystosphenes[112]) or polygonally arranged fissures in the frozen ground[113] (Ger. *Gangeneis*), especially on low plains, e.g. between the Jenisei and Cape Barrow—the cracks, being planes of weakness, open each year so that they become wider and wider. Leffingwell,[114] who stressed this fissure origin, maintained that in Alaska the ice builds vertical wedges in cracks opened by winter contraction, the wedges widening downwards by re-opening and refilling. Similar frost cracks have been seen in Siberia and in Baffin Bay.[115]

Ground-ice may have originated too by burying drift-ice on rivers or lakes,[116], as fluviogenetic deposits in deltas,[117] or when tongues of ice, thrust into beaches, were buried by river-muds and drifting sand.[118] It may have been associated too with heavy deposits of ice congealed over arctic flood-plains,[119] as in Greenland, Alaska, North America, central Asia and north Siberia. The flood-waters, with the growth of anchor-ice and the narrowing of the channel, become restricted, and the impeded flow creates hydrostatic pressure which forces the water to permeate the porous alluvium and force its way out. They also, when frozen over, bulge up and fracture the coating

of ice in weak places by their hydraulic pressure. The cracks freeze and the ice again bursts, flooding and freezing alternating throughout the winter. This "flood ice" (*Aufeis* of Toll and Middendorf, *Näledj* or *Taryn* of Russian geologists[120]) may reach considerable thicknesses, e.g. 3·5 m in Greenland, and sheet a whole plain, persisting over one summer or indefinitely. If covered by fluviatile deposits, it may generate the crystocrene type of Tyrrell and north Siberia's ground-ice,[121] as its structure and association with river mouths suggest.

Ground-ice is usually believed to be a "fossil ice" from the Glacial period[122] though some refer it to the Miocene[123]—this early date is confessedly unsubstantiated and impossible to reconcile with the Pleistocene age of the associated animals (see p. 648). It is a relic of glaciers or firnfields,[124] of flood-ice,[125] of wind-driven snows preserved in the lee of projections, or of snow-drifts[126] or "snow-drift glaciers",[127] formed when the climate was continental[128] or moist[129] or had cooler summers.[130] Its granular structure and high air-content may confirm this age,[131] its persistence being attributed to the protective sheet of "muck", muds, vegetation and soil, and to the fact that since its formation the mean annual temperature has not been above −2°C.[132] It may, indeed, have thinned and in places have melted away completely.[133]

That it represents Pleistocene ice of any kind is denied[134] since it lacks true glacier structure, is unconnected with ground-moraine or other signs of glaciation (Toll[135] thought these had subsequently been removed), and is forming under present conditions (see above). It is, indeed, affirmed that the ice in Alaska is not more than 500–1000 years old,[136] and in Spitsbergen, where it accompanies a recent reindeer head, is not earlier than the close of the postglacial climatic optimum.[137]

This summary betrays only too clearly the uncertainty and confusion that still envelop the question of the age and origin of ground-ice.

Another form of ground-ice is the *pipkrake*[138] (needle-ice; Swed. *pipkrake*; Ger. *Kammeis*; Finn. *rouste*) which occurs as close bundles of fine needles (*Nadeleis*) perpendicular to the cooling surface in fine-grained soils, e.g. clay, marl, loess, sand or peat. It is typically a short-term, e.g. daily, appearance, and therefore forms at or just beneath the surface of the ground in an oceanic subpolar climate, e.g. in Iceland and the sub-antarctic islands, in the German Mittelgebirge, in the tropics, e.g. Mount Kenya and Cordillera Real, and in Japan and the Drakensbergs, and much less commonly in arctic latitudes.

3. *Solifluxion*

The permafrost is important physically and biologically. By the maintenance of much ponded water, it retards the processes of erosion so that the maturing of drainage systems is slowed down even in comparatively friable materials.[139] It has an influence on vegetation,[140] e.g. by retarding the development processes, and by facilitating the formation of ponds increases the richness of the fauna, since the ponds permit many aquatic animals to complete their life cycle before the waters dry up.[141] It also influences the distribution of animals,[142] as in the case of the amphibia which are unable to hibernate in it, of foxes which can have holes only in dry elevated places, and of the earthworms, though these have been found on Kolguev, Novaya Zemlya and the northern coast of Siberia.

The permafrost, by providing an impervious base, gives rise on lower terrain to large swamps or lakes of varying shapes and sizes so that, for example, half of north Canada consists roughly of water and not of land.[143] By preventing percolation and keeping the ground cool and moist in summer, and by developing extensive swamps, it favours the growth of mosses and xerophilous plants.[144] Although it generally hinders chemical weathering[145] (see p. 559), it alternately freezes and thaws the ground[146] and helps to freeze rivers in winter and to build up flood-ice. Geologically, it is most important in giving rise to frost thrust (Swed. *Fjälskjutning*) and frost heave (Swed. *Fjällyftning*), as in the North-West Territories of Canada,[147] in producing soil flow or solifluxion, as J. G. Andersson[148] styled it, and in creating various topographical features,[149] e.g. solifluxion slopes and terraces, soil streams and polygonal marking. This mass movement is a potent planing agent in high latitudes.[150]

Nature. In polar regions, if the ground is not solid rock nor perpetually covered with snow or ice, the climate results in a peculiar type of soil. During winter, the soil is frozen and ice in the interstices wedges the particles apart so that they assume a position of minimum density packing. When thaw comes and water flows freely, the soil and subsoil move like a liquid because each particle is cushioned by a film of water. This soil creeps slowly (the annual rate may be a few decimetres or centimetres[151]) but boldly down the slopes as a semi-liquid or pasty sludge of rock and mud: boulders tend to be aligned in the direction of movement.[152] Its sharply marked boundary with the frozen ground is uneven and itself grades upward into talus and builds masses which slope asymptotically against the hillsides.[153] The bulk of the moving mass is usually fine debris, but as its finer material is being constantly removed it tends to consist of coarse fragments and in this respect to resemble the *Felsenmeer*[154] (see below).

Solifluxion, which may take place on slopes of as little or two or three degrees, is affected by the gradient and structure of the ground, the depth of the thawed layer, the nature of the vegetation, the incidence of the precipitation, and the daily change of temperature. It requires saturation either with repeated freezing and thawing[155] or without regelation,[156] especially if the precipitation is abundant, as in Bear Island, Iceland and the Antarctic islands. It occurs under melting snowfields and where the waters are poorly electrolytic.[157]

Geological action. Solifluxion, one of the factors responsible for the lack of a continuous mantle of vegetation in the *regio alpinum*[158]—it has a marked effect on plants—is an important agent in sorting and transporting soil waste. With the immense flushing and rapid run-off during the spring melting, it may quickly assist in effacing inequalities (cf. p. 1285), such as strandline beaches,[159] in smoothing and rounding hillsides[160] and in making schotterfields.[161] It mixes up fossils from different horizons,[162] transports erratics, as in the Pleistocene Vosges,[163] and makes them round, as in north-west Greenland,[164] or in the case of limestone pebbles of 4–6 cm diameter makes them very flattened and asymmetrical[165]—morphometric analysis distinguisnes them from those of fluviatile and glacial deposits. It shaves off the bedrock, folds and contorts the surface layers to a depth of 7 m,[166] orientates the flatter boulders parallel with the flow,[167] causes terminal curvature directed downhill, and produces chattermarks, polishing and

striations (see p. 246). To it have been attributed flowage and fracturing of rocks, with tilting, step-faulting and trough faulting, to a depth of 30 m or more on valley slopes.[168] Solifluxion also gives rise to asymmetrical valleys (see p. 533) and builds continuous sheets or turf or stone-backed terraces,[169] one or two metres broad, which override the vegetation in front of them and simulate shore features or moraines (they have been falsely interpreted as such[170]). On inclines and morainic hummocks the ground has the appearance of a ploughed field, the furrows running down the slope. Because of the effectiveness of the process, streams above the timber-line in high mountains are ordinarily unable to carry away the detritus as quickly as it is supplied to them.

Solifluxion, of the type termed *Hangsolifluktion* or "clinotropic solifluxion",[171] produces "solifluxion slopes" and continuous aprons which skirt the base of pronounced features,[172] e.g. in ice-free Spitsbergen, Barents Island and Edge Island. While it may sharpen these features, as in the grit and shale country of the south Pennine Chain,[173] and produce altiplanation terraces,[174] as in modern Alaska and Siberia or in Pleistocene Devon and Brittany, it generally grades the sides of the valleys and widens them,[175] e.g. in Scandinavia, Spitsbergen, Greenland and the Falkland Islands, and builds out their floors, overwhelming and clogging the streams, and forming shallow ponds or lakelets[176]; streams have later carved these accumulations into terraces, e.g. in the German Mittelgebirge and the north-eastern Alps.[177] Solifluxion has also produced the dells or "corrasion niches",[178] e.g. in the Mittelgebirge, and here and in Tibet and on Mount Washington the smooth forms of the plateaux.[179] In Spitsbergen, where on inclines exceeding *c.* 15° the slopes are channelled by the stream-formed *Runsen*,[180] it has made some of the U-shaped valleys.[181] It may have carried away all the morainic cover from hillsides in Labrador[182] and have effaced moraines in Spitsbergen.[183] It produces therefore the *Solifluidal* or *Soligelid Formenschatz*.[184] The processes termed *equiplanation*, *altiplanation*, *cryoplanation* or *subnival denudation* are facilitated by its aid[185] and work through a recognisable cycle.[186]

Yet another product of solifluxion is the "stone-river". Observed in Tibet by S. Hedin, it characterises the Falkland Islands[187] where it was described by C. Darwin (1846) and C. Wyville-Thomson (1877) and more fully by J. G. Andersson (1904), and interpreted by L. Agassiz as a ground-moraine.[188] These accumulations of moss-grown boulders of quartzite and sandstone, which are up to 6 m long and 40 tons in weight and occupy the valleys, were formed by the excessive freezing and thawing incidental to the glacial border conditions of the Ice Age and the present climate.[189] They have been recently attributed primarily to frost acting on quartzite faces and only subordinately to creep or solifluxion.[190]

Rock-glaciers, "rock-streams", "rock-rivers" or "crystocrenes" (Fr. *glaciers de pierres, coulées de blocs*), which skirt the foot of steep rock-faces in the cirques of Alaska[191] and of western North and South America in lower latitudes[192] and sporadically in the French Alps,[193] the Engadine,[194] Tyrol,[195] Carpathians,[196] Urals[197] and Greenland,[198] are glacier-like piles of mud and coarse angular waste which exhibit convex, tongue-like forms and are corrugated into ridges parallel with their sides and front. Their surface slopes at an average of 9–18°. They contain interstitial ice, particularly in the lower layers, and move in some such way as a glacier, advancing at varying rates and

displaying hollows and ridges not unlike those of moraines. They may indeed be regarded as dying or "fossil" glaciers,[199] covered with end-moraine and ablation moraine. They may, however, be merely landslides,[200] and owe their ice to the freezing of springs entering them from the hills.[201] Thus photographs taken 70 years apart yield no evidence of such movement in the Presidential Range.[202]

Solifluxion also produces stony clays or pseudo-tills (see p. 1075).

Distribution. Solifluxion, as surveys of its distribution show,[203] is bounded by definite climatic limits which, like the snowline, rise towards the equator and from the coast inwards: the lower limit, for example, is at 1000 m in north Scandinavia, at 1500 m in the Riesengebirge, at 1800–1900 m in the northern and eastern Alps, and at 2000 m in the central Alps and Balkans. It characterises polar and sub-polar lands and high mountains in lower latitudes; it has been reported from the Arctic,[204] e.g. Alaska, Greenland, Iceland, Spitsbergen, Bear Island, Novaya Zemlya and north Canada (it is general over much of the subpolar region), from the Urals,[205] Fennoscandia[206] and a few Alpine localities[207] (the permafrost is generally absent), and from central Kurdistan,[208] Kerguelen[209] and Macquarie Island.[210]

4. *Polygonal Markings*

Character and classification. Solifluxion gives rise on steep declivities to parallel ridges or "striped ground" or "stone stripes" (Ger. *Streifboden*; Fr. *sols rayés, sols rubanées*; Norw. *rutemark*), up to 2 m wide, as for instance in east Greenland[211] and the Antarctic.[212] The size and spacing of stone stripes depends largely on the degree of slope and the size of the fragments. On steep slopes lateral movement is small and the stripes are close together, and on slopes much over 30° the movement is almost wholly downhill and stripes do not form. Transition forms[213] link these "free types"[214] with "polygonal markings" or "fixed types" which replace them on gently rolling or nearly horizontal ground. Though geologically unimportant, their appearance is striking and is enhanced by plants which grow in the channels or in the centre of the polygons.[215]

The polygonal or patterned ground[216] (Ger. *Polygonboden*; *Rautenboden*, *Karreeboden*; Fr. *sols polygonaux*; Swed. *rutmarken*), since first discovered over one hundred years ago and more particularly since the excursion of the Stockholm International Geological Congress to Spitsbergen in 1910, has been much discussed and often classified.[217] Högbom's classification,[218] which has found some adherents,[219] recognises the following types which by continual comminution of their material may represent an evolutionary series[220]: (1) stone-nets, (2) stone-rings, (3) mud-flat polygons (*Schuttinseln*), and (4) fissure-polygons (*Zellenboden*). There are numerous synonyms.[221]

Stone-nets (*Strukturboden* of Meinardus, stone polygons of J. G. Huxley and N. E. Odell) build contiguous polygons (pl. XXIIIB) of varying diameters[222] but usually 0·5–2 m (Norway, 3–5 m; Iceland, 1–1·5 m; Greenland, 4·5 m; Himalayas, up to 1·5 m; Antarctic up to 15 m), the diameters depending apparently upon the suitability of the ground and topography and upon the amount of water in the soil.[223] The diameter in Iceland increases with latitude.[224] The shapes are occasionally elliptical or circular[225] and are pulled out in the direction of slope if this is steep so that finally on a slope of

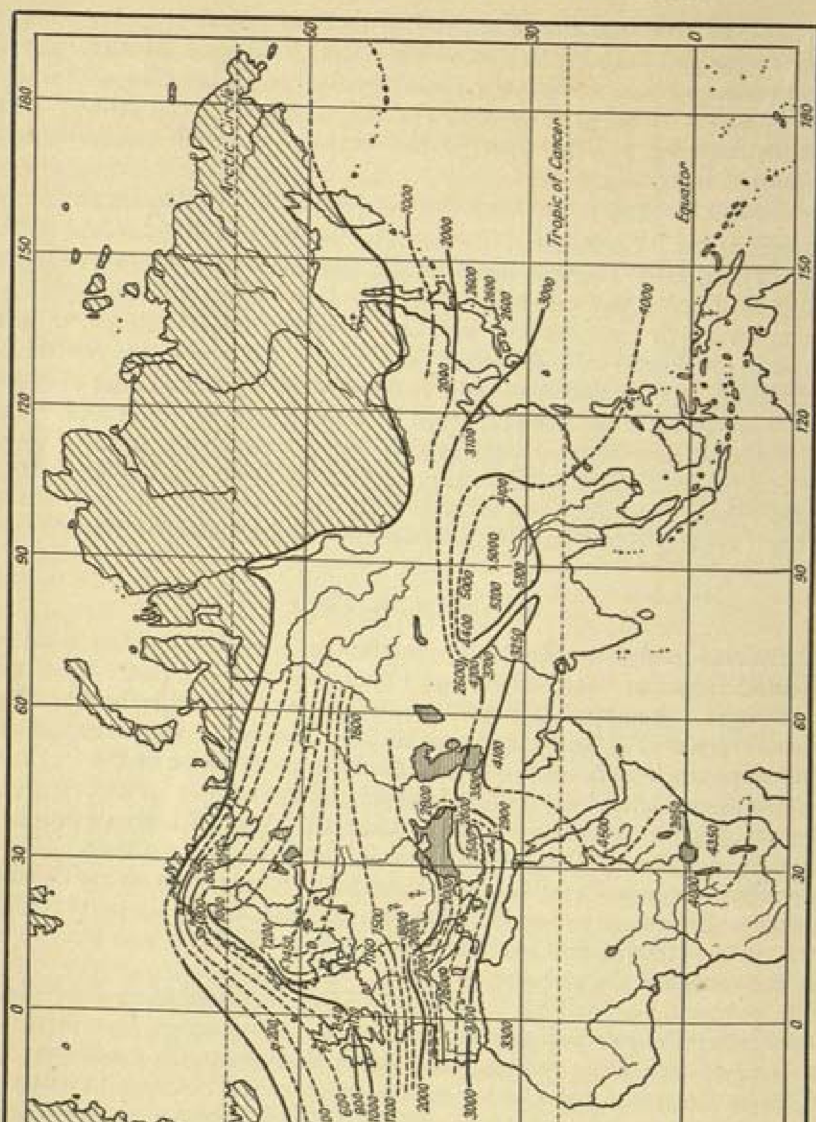

FIG. 111.—Distribution of the present *Strukturboden* (altitude lines) and permafrost (shaded) in the Old World. C. Troll, *1718*, p. 164, fig. 1.

about 5° the sides become parallel and follow the steepest pitch. Their material is heterogeneous. The stonier element, of various sizes, is restricted to the edges, and if platy assumes a vertical position. The centres which are usually convex upward (they are occasionally concave and filled with smaller pieces[226]) and clothed with moss or similar vegetation are sometimes cracked radially[227] or subdivided into secondary or tertiary polygons.[228]

Excavations prove that the features are shallow[229] (the average depth is

0·5 m); experiment shows that the depth is proportional to the diameter[230] and mathematics that it equals one-third or one-fourth of the diameter.[231]

Stone-polygons are chiefly seen on water-logged terrain, such as hard rock,[232] ground-ice,[233] raised beaches, or the flatter parts of moraines, notably if their material is angular or mixed.[234] They are absent from steep slopes and are rare on perfectly level ground—the critical angle is 5°.[235] For these reasons—also because any flat surfaces at these altitudes have firn or if bare suffer from strong insolation—they are few on the high Alps.[236] If they occur singly instead of in their customary groupings they constitute Högbom's "stone-rings"; transitions reveal a close connexion between the two types.[237]

Mudflat polygons,[238] like the following type, are confined to homogeneous materials, especially muds, alluvium, deltas, humus or peat. Fissure or mud polygons grade into stone-polygons by the incoming of stone-borders (pl. XXIIIA and B). Bounded by cracks, they range up to 20 m in Spitsbergen[239] or 30 m in Siberia.[240] According to Odell,[241] they are made by winter frost and may descend from stone-polygons by breaking down the stone-borders; their secondary polygons are due to drying. "Arcuate markings", a type which Huxley and Odell established, are only seen in shales.

To the largest form of the polygonal ground belong the "tundra polygons", "Taimyr polygon", "tetragonal ground" or "chess-board ground".[242] These forms, polygonal or tetragonal in shape, show no sorting and are bounded by ice-filled cracks and earth-walls above the cracks. They occur in Siberia, Spitsbergen and Greenland.[243]

Forms which also show a regular relief pattern and are associated not merely with frost but also with biological processes are those which occur in grass (*Rasenböden*, etc.), peat (*Palsen*), and other kinds of vegetation.[244]

Origin. The problem of the origin and development of polygonal ground will probably only be solved when thorough investigations of a physical and experimental nature are made, together with local observations on temperature, freezing, water circulation, etc. The weakness of almost all pertinent hypotheses is that they are based on observations of external properties alone. The features are merely superficial structures, restricted in depth to the *Auftauboden* and the limit of freezing.[245] They are generally, albeit by no means universally,[246] thought to be connected with frozen ground, which keeps the temperature low and the ground moist at the thaw, prevents percolation and makes the ground above it into a "closed system". Permafrost may however be partly replaced by impermeable rock close beneath the surface.[247] Thus polygonal ground occurs in Iceland where there is no permafrost.

Notwithstanding this amount of agreement, there are almost as many views as observers; for probably about thirty hypotheses have been suggested, many of which contribute to a full explanation, though this does not apply to theories which invoke earthquakes,[248] mud-eruptions,[249] ice-rafts[250] or thermal or CO_2 springs.[251] Some writers as we have just seen find the link in the lower limit fixed by the frozen ground to the penetration of water; others find it in the gliding surface the permafrost provides. Whatever the cause, it operates rapidly since polygonal markings appeared in Lapland on ground a swollen river had built only three months previously.[252] Finality will only be reached when more experiments have been made on porosity, degree of compactness, ability to flow, and the susceptibility to frost along lines already begun.[253]

The markings are probably due to repeated freezing and thawing in the top layers.[254] Frost pulverises the rock-fragments of the centre[255] or sorts them vertically (frost heaving[256]), by differential heaving and gravity settling, and horizontally (frost thrusting[257]), by pushing the coarser outwards to leave the finer ones in the middle—the doming of the centre is due to heaving, to distension by more abundant water in the fine material, to the increase of fine material by comminution, and to the formation of thin lenses of ground-ice. Field and laboratory studies indicate that most disruptive effects of freezing are due to the growth of ice-crystals rather than to change in volume.[258] Fine material is also drawn inwards by cohesion.[259] If the stones are flat, the upward pressure naturally pushes them upright. In this way fossils are raised out of their original horizon.[260] It is suggested (*a*) that frost forces the mud and stone outwards but on contraction the stones tend to stay where they are and the muds to be drawn back to the original position by the cohesion of their particles[261]; (*b*) that frost differentiates the material by splitting the stones in slight hollows which are damper than the ground around[262]; (*c*) that the process of splitting and comminution, once the stones and mud have become differentially arranged, are self-reinforcing in the mud-centres but retarded in the stone-borders[263] where gutters in the frozen ground serve as drainage channels and remove the finer material from the borders and inhibit the comminution of the stones in the upper part[264]; and (*d*) that the severer frost in the lower, sodden layers produces the finer material and brings it to the surface, pushing the coarser to the sides. The stone-rings so produced grow outwards from their centres to build a polygonal network.[265] The activity is probably highest near the surface and diminishes towards the bottom of the thawed soil since the polygon cores taper downwards.[266]

How the centres themselves came to be fixed is omitted from most theories. They may, however, be located by the behaviour of the snow upon an uneven surface[267] or be domed by water rising in them[268] because of evaporation. Alternatively, they may initially have lain within hexagonal cracks formed by contraction[269] at sub-freezing temperatures[270] or have been connected with convection streams on the lines of H. Benard's classic experiments[271] and the calculations of Lord Raleigh[272] and A. R. Low[273] (see below), or have been areas where drainage was slow and mud collected when differential comminution began. The amount of water in the ground is a factor,[274] as is the size of the material.[275] Once the centres were full enough of mud, they became centres of freezing and thawing and convex because of the process of upwelling.[276]

Some workers,[277] relying partly on good experimental support,[278] relate the polygons to circular movements in a vertical plane (*Brodelhypothese*) as was first suggested by O. Nordenskiöld (1907) and later developed by K. Gripp. Frost as the driving force is replaced by convection currents connected with differences in density at various levels, controlled by temperature distributions (water is densest at 4°C) and the selective absorptions of the sun's heat or the increasing content of water towards the permafrost. But this, it is said, is unlikely, for the bringing up of fossils and rocks from the lower layers and the vertical slabs which have been mentioned as evidence, are not crucial—they are as readily explained by the alternative theory of frost heaving and thrust. Moreover, the temperature fall of 4°C from the surface to the frozen ground, as has been shown by measurements in Spitsbergen and

the Alps, does not take place,[279] and the convection, which is contrary to fundamental laws of physics, is negligible in these shallow depths[280] and would not convey differently sized boulders at the same rate. The stones in the mud polygons are not arranged in the way that convection would arrange them and those in the borders are at the surface and have not been carried down again. Yet a convection due to freezing above an unfrozen layer, the winter's "aspiration" of water and a spring ascent of clay is possible[281] and accords with S. Taber's observations.[282] Colloidal processes have also been invoked,[283] especially in clays and humus, in tropical forms and for the remarkable features seen in the sand plains on Jan Mayen,[284] as well as processes of hydration and dehydration (Schenk, 1955).

Fissure polygons have been linked with desiccation[285] but, except in early stages when they are often associated with damp ground, are probably a product of frost and relief from frost tension,[286] though ground drying in summer may promote their formation: the differences between polygonal ground and desiccation features have recently been tabulated.[287]

Polygonal markings probably result from alternate freezing and thawing, from solifluxion, from the swelling, shrinkage and flowing of clays, and from the establishment of vegetation with its attendant insolation and filter effects.

Frost scars, peat rings, tussock rings, tussock groups and tussock birch-heath polygons form an evolutionary series in Alaska in the silting of mineral soils beneath peats.[288]

Distribution. Polygonal markings were noted in Novaya Zemlya by K. E. v. Baer[289] in 1837. They have since been found widely distributed in polar lands where the severe cold, resulting from the low angle of the sun's rays and their long traverse of the atmosphere, the strong cloudiness and mistiness of summer, the melting snows of spring and summer, and the presence of ice in the ground are conducive. The polygonal ground is found also where the ground is frozen or the water circulation is very shallow. It occurs in heterogeneous matter like fans, moraines, alluvium, river-terraces, or raised beaches which are free from snow for part of the year. It is occasionally seen in standing water.[290]

In the Arctic,[291] e.g. Spitsbergen and Iceland, it extends down to sea-level and in the Faeroes[292] to 200 m, in Torne Lappmark to 800 m, and elsewhere in central Scandinavia to 1000 m[293] and in south Norway (Jotunheim) to 1450.[294] The climatic limit (Ger. *Strukturbodengrenze*) lies near the snow-line or several hundred or even thousand metres below the snowline—it has therefore been called the "arctic" (H. Poser, W. Salomon), "polar" (S. Passarge) or better still the "subnival" form (C. Troll). Like this line it ascends to considerable heights in lower latitudes[295] (fig. 111), e.g. 1400–1600 m in the Riesengebirge, 1800–2200 m in the Alps, 1900–2000 m in the Rila Mountains, 2150 m in the Caucasus, and 4500 m in Kibo and the Cordillera Real.[296] In the Iberian Peninsula it rises from *c.* 2600 m to *c.* 3000 m in the Sierra Nevadas and to 3250 m in the Punta de Travelez.

The highest altitudes are reached in the subtropical pressure belts, e.g. in Tibet (5200 m) and the Andes (4700 m). It also rises from the oceanic lands into the continental areas, e.g. Faeroes 200 m, Scotland 600–800 m, Riesengebirge 1500 m, Central Alps 2000 m, west Pamirs 2600–2800 m, Lebanon 2900 m, east Pamirs 4700–5000 m and on Mount Everest 5000 m, from the lowlands on Sakhalin to over 2500 m in Manchuria and

c. 3000 m in north Shansi, and from 350 m on Mount Desert Island, Maine, to 1500 m on Mount Washington.

The markings have been discovered in the Arctic[297] (Baffin Land, Jan Mayen, Greenland, Iceland, Faeroe Islands, Spitsbergen, Bear Island, Franz Josef Land, Novaya Zemlya); in temperate latitudes in North America[298] (e.g. Mount Washington, Presidential Range, Mount Katahdin, Mount Monadnock, Mount Desert Island, Gaspé Peninsula), Hawaii,[299] Europe[300] (Fennoscandia, Vosges, German Mittelgebirge, north Germany, Alps, Pyrenees, Great Britain), north Africa[301] (High Atlas), Asia[302] (Lebanon, Taurus, Pamirs, Himalayas, Siberia, Mongolia, Japan), and the tropics[303] (Kilimanjaro, Mount Kenya, South American, Cordillera), New Zealand[304] and in the Antarctic,[305] including the Antarctic islands. The tropical and subtropical type, which is related to daily oscillations, and that which is found in sub-Antarctic islands (Kerguelen, South Georgia) and in coastal Iceland, is of the same origin as the polar type which is connected with seasonal changes and also occurs in Scandinavia, Riesengebirge and the Ural Mountains but in miniature[306] (10–25 cm).

Many of the occurrences are quite recent,[307] including those which have been found on exposed lake-floors in Rhätikon, within the 1850-moraines of the Alps, and on quarry spoil heaps near Loch Lomond. The markings have been observed in a depth of 1–2 m of water off west Greenland.[308]

5. *Blockfields*

Character. The flat terrain at higher elevations in middle and higher latitudes is often covered with a sea of angular blocks variously called "block fields", "boulder fields" (Ger. *Felsenmeer*, *Blockmeer*, *Blockfelder*). Composed almost wholly of local rocks, this mountain top frost-debris may be a few metres deep and may completely bury the subjacent rocks, notably if these are quartzite, sandstone, limestone, slate or massive igneous rock, e.g. granite or basalt. The blocks often diminish progressively from the summit towards the margin where they merge downhill into genuine talus. Transition forms glide or creep down the slopes along depressions[309] and in parallel waves or steep-fronted and flat-topped terraces in which fragments are upheaved perpendicularly to the slope of the front and shrink vertically under the influence of gravity. Frost-riven material forms, for example in Scandinavia,[310] a zone quite distinct from the "Alpine zone"—it has been termed the *regio alpina* II, *regio alpina secunda* or *regio alpina sterilis*—and is either barren of life or has small patches of moss, lichens and *Salix herbacea* and *Ranunculus glacialis* associations. It may even underlie plains in the Arctic, e.g. the Siberian stone-tundras[311] (pl. XXIVA, facing p. 571).

Origin. Blockfields which are found in polar latitudes,[312] e.g. Kerguelen, Greenland, Alaska, Iceland and north Fennoscandia, and in more southerly climes (see p. 1199), have been linked with the removal of the finer material for the drift[313] or solifluxion soils.[314] They were, however, disintegrated mechanically by frost acting along structural planes in hard, well-defined types and horizons of rock[315]: the finer material has been blown or washed away. Disintegration is most rapid[316] if there is no vegetation and the climate is severe, if there are ample moisture-laden winds—it diminishes towards the colder and drier areas of the far north[317]—or if the temperature changes with foehn winds, as in present-day Greenland[318] where the winter

temperature may rise from $-30°C$ to $-10°C$ in a few hours. It is most severe near the snowline[319] on concave surfaces, e.g. in the Auvergne and German Mittelgebirge, and generally on northern and eastern faces in the northern hemisphere.

The number of oscillations through freezing point rather than the degree of cold is the controlling factor[320]; the vertical distribution of such oscillations has been investigated for Switzerland[321] and the German Mittelgebirge.[322] The *Frostwechseltage*[323] (of air temperatures) are about 57 in Spitsbergen, 98 at Snow Hill, Graham Land, 120 in Kerguelen, and 79 on Schneekoppe (Riesengebirge). S. S. Visher[324] has mapped them for the United States (fig. 112).

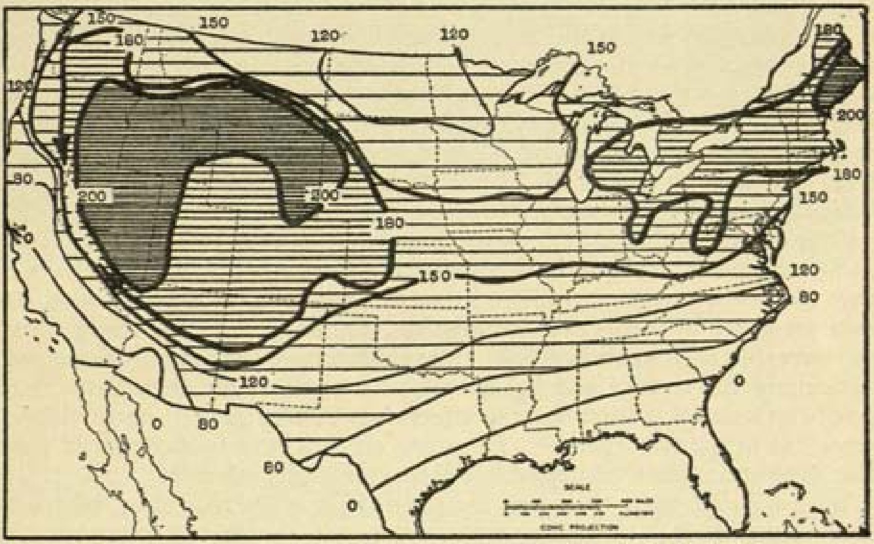

FIG. 112.—Average annual number of times of freeze and thaw in the United States. S. S. Visher, *G. S. A. B.* 56, 1945, p. 731, fig. 24.

Frost acts only slightly in ground which is more or less permanently frozen[325]; this is true in the Antarctic[326] and is implied by the rounded rock-shapes on its central plateau as Amundsen noticed. Exceptions are found at the continental margin[327] where periods of daily freezing and thawing mark the two ends of the winter season and may be pronounced because the daily insolation and radiation are higher. The temperature changes in lower latitudes with day and night but in higher latitudes, where insolation effects are marked and the sun's course more or less parallels the horizon, it results from sun and shade[328] which relief or drifting clouds induce.

The material in the *Blockmeer* has in places moved considerably. Frost and ice-crystals in the ground, water liberated by melting snows, and gravity acting on rocks subject to changes of temperature have provided the actual means.[329] The frost works as a levelling agency; it may have produced, for example, the lawns, benches, spurs and tablelands of the Mount Washington Range.[330]

Nunataks and blockfields. A nunatak (*nuna*, lonely; *tak*, a peak)—"nunataq" is a variant spelling[331]—is an Eskimo term for an island of rock

or mountain peak in a sea of land-ice; A. E. Nordenskiöld introduced the word into glacial literature. "Marginal nunataks", like the Jensen and Dalager Nunataks of west Greenland,[332] have ice on three sides only and on the fourth are bounded by sea, fjord or land. Polar Eskimos use *pingo* for a mountain which is entirely hidden by ice but sets its mark upon this.[333] "Nunakol",[334] a compound word coined for nunataks rounded by ice, is objectionable etymologically and because such forms are difficult to distinguish in the field.[335]

That the ice once rose high up the sides of nunataks (many of them "monadnock-nunataks") is proved by "tails" of debris or solid rock, by moraines or lines of boulders which stretch from their lee, or by lateral moraines whose successive rings record the nunatak's stepwise emergence.[336] Theoretically, the height is given by the highest erratics and glaciated surfaces or by orographic features, including stoss and lee slopes.[337] But to apply these criteria is difficult since an angular, serrated peak may be a true nunatak or a moulded summit deprived lateglacially of its glaciated forms[338]. The effacement may have been by avalanches, as in the North Limestone Alps,[339] by solifluxion, as in Spitsbergen,[340] by desquamation arising from insolation, as on Gaussberg,[341] or by frost which was more harmful to rocks than to unconsolidated accumulations.[342]

Erratics rest on the rugged summit of Mount Schurmann Nunatak of the Cornell Glacier[343] and granite veins in east Greenland have been lowered 5 m since the ice overrode them.[344] In west Greenland,[345] glacial polish has been preserved in places even on granites, e.g. under erratics, though frost may have split off the polished slab. But erratics have been usually destroyed, particularly the smaller and lighter coloured ones, as on Gaussberg,[346] or have been reduced to frost-riven splinters, as in Spitsbergen,[347] or to "debris-cones" as in the Antarctic.[348] Balancing erratics have been removed from their base and roches moutonnées broken up in Kerguelen.[349]

Angular relief may, therefore, be compatible with overriding by ice[350] since, at the levels of the higher peaks, the ice was thin and clean and moved slowly and frost acted drastically. The difficulty of distinguishing between true nunataks and overridden peaks is seen in the contrasted views that are held concerning the glacial history of such high marginal lands as the Lofoten Islands (see p. 1302), parts of west Greenland (see p. 850), north British Columbia (see p. 853), Shickshock Mountains (see p. 853), and the Torngat Mountains of Labrador (see p. 853), where high humidity and great cold encourage frost action.

6. *Lake-ramparts*

Character and distribution. Lake-ramparts are walls of boulders or other materials which have their base above and beyond the ordinary lake-shore and follow its inflections (pl. XXA); their composition and irregular plan and crest readily distinguish them from ordinary lake-beaches.

These features—alternative names are "ice shove ridges", "ice push ridges" or "ice pushed ridges"—have often been observed[351] since attention was first attracted to them in North America in the early part of last century.[352] E. R. Buckley,[353] in a particularly full description, recognised the following types: (1) ridges of sand and gravel on gradually sloping beaches; (2) irregular blocks piled up into ridges or thrust into clays on abrupt shores, the

A. Lake-rampart, Sucker Lake, Shellbrook, Saskatchewan [W. O. Kupsch]

B. Blockfield on Mount Ikordlearsuk, Torngat Mountains, formerly glaciated [N. E. Odell]

(facing p. 576)

A. Mud polygons near Brucebyen, Billefjorden, Vestspitsbergen
[H. B. Harland]

B. Stone polygons, Kapp Ekholm, Billefjorden, Vestspitsbergen
[H. B. Harland]

(*facing p.* 577)

cliff being fractured and its brow, if not high, pushed over into a rampart; and (3) folds, with thrusts and tear faults, on low marshy ground about the heads of bays. If the shore is of such a nature that it will not yield, the ice buckles with no effect.

Ramparts are found to-day in higher latitudes, e.g. along the Mackenzie River,[354] in Vermont and Wisconsin,[355] in Denmark, Germany and the Baltic States,[356] in Scandinavia,[357] in Greenland,[358] in central and northern Asia[359] and in New Zealand.[360]

Origin. Shore-ramparts require sudden oscillations of air-temperatures over a frozen lake. A rise of temperature expands the ice, induces stresses of considerable strength,[361] and crowds the edge upon the shore (the advance on Minnesota lakes was 6 in. (*c.* 15 mm) per diem[362]), thrusting up the shore-material, uprooting trees, and damaging or overturning piers or sea-walls. Fissures open by a collapse of the ice following a lowering of the water-surface[363] consequent upon a fall of temperature—ice is weak in tension—the cracks appearing to the accompaniment of loud reports audible over many miles. The cracks, which may form a set radiating from the centre of the lake and another set roughly concentric with the shore, fill immediately with water which on freezing extends the ice. For big effects, the various processes must be repeated numerous times.

Normally, the ice has enough weak points to permit the release of the pressure by buckling the sheet or by crumbling its edges. Moreover, a cover of snow hinders the happenings by preventing the temperature changes from being communicated quickly to the ice and by bending the ice and leading to buckling and loss of horizontal energy. Furthermore, deep lakes which do not have a heavy coating of ice are without ramparts, as are large lakes except in their smaller bays, since either the ice is incompetent to constitute a strut and buckles into pressure ridges, or storm waves in summer obliterate any ramparts that may be produced.[364] Observations suggest the limiting diameter is *c.* 800–3000 m.[365]

The thrust acts to a distance which depends upon the size of the lake and the local climate. Hence each lake has a definite limit. On shelving shores the ice is forced up, together with any loose boulders around which it may freeze. Small stones are carried offshore in summer, big boulders in winter; the successive operations yield a constant selection.

The expansion is aided, in some cases not inconsiderably, by wind working on loose ice. Suitable winds on big lakes during a spring melting blow ice-rafts across the shore-lead and against the shore with great violence, as on modern lakes[366] and Pleistocene Lake Maumee.[367] J. B. Tyrrell,[368] from observations in Canada, advocated this view. In his opinion, the ice which thickens mostly from the top by the freezing of the overflowing waters, and in areas of light snowfall mostly from the bottom, expands laterally with the fall of temperature though most of the expansion on warming is vertical. Throughout the winter, the ice remains immovable and firmly frozen to the shore. At the spring thaw, the ice first melts along the shore to form a lane surrounding the large central mass. Ramparts are formed by this loose, floating ice which winds drive this way and that against the shore with prodigious force. This is without doubt a very important factor, especially on large lakes: it changes or destroys sand bars, causes recession of banks, and shifts the position of boulders.

Geological action. Lake-ice acts geologically[369] by transporting material from the lake-bottom to the beach. It may also, by breaking the shore, enlarge lakes and shallow the lake-basin. Sometimes, as on Lake Windermere and in Finland and Canada, the ice striates and grooves the rocks, the markings being strictly limited in length and perpendicular to the shore.

7. *River-ice*

The topographical effects and cumulative geomorphological results of river-ice are only slight. In northward-flowing rivers, e.g. in Alaska, Canada and Siberia, it gives rise at the spring melting to ice-runs, ice-jams and devastating floods.[370] Like lake-ice, it shears the surface earth and rocks over the unfrozen subsoil, damages piers and walls, erects ramparts along its banks,[371] as along the north Siberian rivers and the Yukon and Mackenzie rivers in North America, and occasionally in Europe, e.g. the Neckar.[372] It furrows the banks and keeps them permanently free of forest by bending and uprooting trees and lacerating them; produces boulder-pavements[373] whose upper surfaces are somewhat abraded; and sometimes abrades and plucks and fashions roches moutonnées.[374] It striates and polishes the rocks[375] parallel with the banks or slightly oblique to them, noticeably on the upstream side of bold promontories, on the outside of river-bends, on rocks dipping upstream, near strong currents, and at the sudden break-up of the ice. The striations differ quantitatively rather than qualitatively from those glaciers engrave but are usually short and coarse, frequently curved and unrelated to well-marked facets. They are naturally less regular and firmly drawn, since river-ice is less intense, continuous or prolonged in its action, and rarely has sharply cut planes to score. At the bursting of lakes held up by ephemeral ice-jams, rivers work with redoubled energy and may cut deep gorges; such, it is suggested, are the multiple channels of the scabland of the Columbian uplands (see p. 240).

River-ice rarely grinds flat facets[376] but it conveys erratics, as observed in recent times[377]—in New Brunswick material has been carried *c.* 800 m and sometimes contrary to the direction of the Pleistocene glaciers.[378] Boulders 3 m long may be so transported.[379] Anchor-ice moves material through short distances only since its spongy and loose texture favours rapid melting.[380]

During winter, water-soaked soils and water standing in depressions above them in flood plains become frozen into one coherent mass. With the spring freshets and the submergence of the plains, the mass of ice with its burden of soil below is buoyed up and floated downstream. In this way, masses of soil, 10–15 cm thick and 15 or more metres across, are transported.[381] Its cobbles, owing to the smaller and less continuous pressure as compared with glacier-ice, have short, irregular and random striae, and are almost exclusively subangular to well rounded, though snubbed ends or edges or facets have occasionally been produced.[382]

Acicular ice-crystals or needles, impinging on fluviatile, lacustrine or marine muds, as well as other signs of ice-action, may leave their impress in recognisable markings.[383] They are the Swedish *pipkrake* (see p. 566).

8. *Avalanches*

Snows which exert a climatic influence wherever they fall have a geomorphological significance only when they are plentiful and lie for consider-

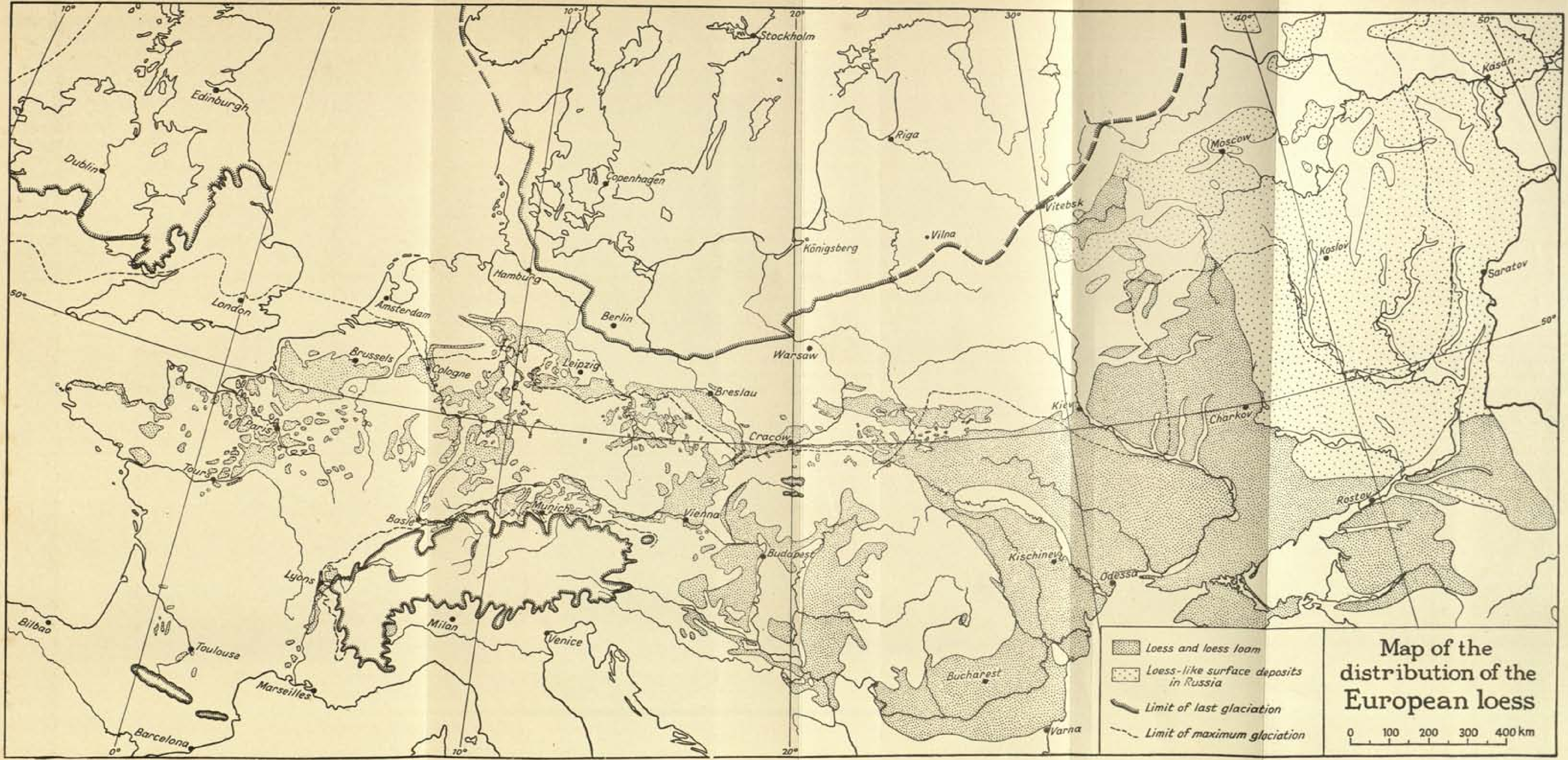

FIG. 101.—Map of the European loess. R. Grahmann, 637, pl. 1 (opp. p. 24).

To face p. 578

able periods; they prevent frost heaving and wind erosion. Yet periglacial snow erodes not a little. It does so when more or less stationary, as by nivation (see p. 302) and by its weight on trees and other vegetation,[384] and when in movement, either as wind-driven clouds, as masses gliding almost imperceptibly over the ground and scratching it,[385] or as avalanches (see p. 19).

Drifting snows at low temperatures, because of their hardness (see p. 214) and snow blast, destroy vegetation, including grass and trees by burying plants, damaging roots and stems, changing soil-water relations and causing instability of soils.[386] They polish and erode rocks[387] in the Arctic and Antarctic; fray rope, etch wood and polish metal[388]; smooth and round the windward faces of rocks and form vermiculations in ice and rock; and raise in marked relief the harder parts of heterogeneous strata. Smoothed, pitted and glazed stones occur in the Antarctic[389] and brown, polished rocks near the Gepatschferner and other Alpine glaciers.[390] The term Arctic Sahara, fittingly bestowed by E. Whymper upon Greenland, recalls the barren wastes, clear skies, high winds and erosive sand blast of the desert.

Yet this action is insignificant compared with the performance of avalanches. These, though intermittent, are excessively destructive. By compressing the air and inducing hurricane-like winds, they may initiate new avalanches and lay low whole woods up to 1 km away, tear off the roofs of buildings, and blow up stone or iron bridges, lifting large fragments 15–45 m into the air.[391] Besides these ravages, they act as one of the most devastating forces in Nature; J. Coaz mapped 9368 important avalanche tracks in Switzerland and P. Mougin 1361 for Savoy. They are particularly powerful in higher valleys, eroding the slopes over which they fall,[392] smoothing the surfaces and scoring parallel chutes (Fr. *cannelures*) in rocky cliffs,[393] and helping to excavate cirques.[394] They tear off grass and vegetation, destroy the surface-layers where they creep slowly,[395] and plough up any forests they traverse. They smooth the floors and gouge out the walls of their gullies,[396] and scratch and polish the rocks in their paths.[397] The striae and grooves are usually short and scored downhill and not along the valleys as in the case of glaciers; those on boulders are restricted to one side.[398]

Avalanches may carry vast quantities of debris which they detach from projections at the sides of ravines and gather in the process of steepening their walls. They help to build the snowslope moraines (see p. 407) or, more commonly, deposit the detritus on coming to rest as cones up to 20 m thick,[399] the material being sorted according to size and grain.[400] In one year (1909–10), the avalanches of Savoy built up cones of debris with a volume of 23,279 cu. m; a single avalanche in Chamonix brought down in one day 2000 cu. m of earth and rock.[401]

Avalanche cones do indirect damage. They pond rivers or streams in temporary lakes[402] which, by suddenly bursting their dams of snow and incoherent debris, send violent floods rushing down the valleys. Ice-avalanches may produce catastrophic floods in the same way.[403] They may also create great waves where they fall into the sea: the Falling Glacier (see p. 111) created waves which near the glacier washed out and broke off alder bushes at 110–115 ft (*c.* 33–35 m) above sea-level, and 3 miles (*c.* 5 km) away rose to 55 ft (*c.* 17 m).[404]

The geological significance of recurring avalanches is therefore considerable; for they are much more widely distributed than glaciers. Their role to-day is on a scale not adequately realised. They were similarly active

during the Pleistocene[405] in the later stages of the retreat[406] as well as during interglacial times,[407] especially while the ground was still bare of vegetation. The oversteepened hillsides favour their action. Many screes were partly built up with their aid; and the floods which attended them strewed the valley floors with blocks.

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CHAPTER XXVIII

GEOLOGICAL ACTION OF SEA-ICE

Detritus in sea-ice. Sea-ice depends for its geological action[1] partly upon its detritus. Although it is generally clean and free from detritus, as Darwin[2] noticed, mud and stones confined to cylindrical or cryoconite holes or disposed irregularly in heaps upon its surface have been observed in the Arctic,[3] as by Sir. J. Franklin, Sir J. Ross, Sir. W. E. Parry, and particularly by E. K. Kane[4] and later explorers. Mud-covered sea-ice on the shelf of Siberia is so common that the ice-fields have a brownish colour[5]—this source is partly responsible for the cryoconite found in the pack-ice north of Spitsbergen[6] and east of Greenland.[7] The Antarctic pack-ice too contains debris,[8] though probably much less than in the Arctic since the ice-sheet and bergs fend off the sea-ice from the land. Transport by drift-ice takes place on the Baltic coasts.[9]

Drift-ice, which may convey huge amounts of shore-debris and rock-fragments up to one-fifteenth of its mass,[10] derives its material by falls[11] from landslips, screes or avalanches from steep hill sides, or by freezing on to beach shingle, sands and shells when driven ashore.[12] Offshore winds[13] and rivers[14] add to it, and the sea casts sands, shells and seaweed upon it.[15] Bergs which get their impurities from the margins and basal layers of their parent glaciers have little material.[16] Antarctic bergs like Antarctic sea-ice are remarkably free from foreign matter[17] (moraine-charged bergs in the Weddell Sea are only 1–2% of the total number) since they calve from shelf-ice which is fairly clean.

Erosive action. While drift-ice on the whole acts but insignificantly on solid rock,[18] it erodes not a little when driven into shoaling water by on-shore winds and gales, and with the weight and force of the entire pack behind it. Grinding bergs and heavy floes tear off algae, an important drifting constituent in the West Greenland Current,[19] and hinder algal growth, and cause an almost complete lack of sessile animals over much of the littoral and sub-littoral zones in both the Arctic and the Antarctic.[20] They score the sea-floor and even it[21]; alter the contour of the bottom, e.g. near Point Barrow[22]; and build shallows, banks and islands,[23] adding millions of tons of material to the seaward side of reefs.[24] They thud like titanic battering-rams against the tundra cliffs, break and shatter the rocks, fold and contort them if these are soft and incoherent,[25] and erect barrier beaches, ramparts and lines of boulders, closely packed, as along the Arctic and Antarctic coasts[26] and in the Baltic Sea.[27] They also round, polish and groove,[28] and leave pits and long, sinuous trails as they drag with the rise and fall of the tide.[29]

While sea-ice because of its structure is perhaps not able to scratch or erode so well as land-ice,[30] it is unquestionably able to striate rocks, as was so generally affirmed during the earlier decades of the last century; for, although many Arctic coasts were the scene of Pleistocene glaciation, carefully sifted evidence proves the action. Striae scored by this agent[31] are usually

irregularly distributed and not certainly distinguishable from those inscribed by glaciers (see below); they are parallel to the shore if engraved by bergs, and perpendicular to it if scored by coast-ice which sometimes inscribes them by tidal rise and fall or by sliding over stones in the water. They are seemingly not so deep or regular as glacial striae; they are often curved and concentric or broken and crossed; and they leave untouched any hollows between projections or unevennesses. Chattermarks do not accompany them.

Bergs erode[32] indirectly when calving or overturning. Their calving floods,[33] which exceed all ocean waves in power and size and may generate waves 30 m in height (at beach projections even 100 m) at the actual scene of the calving and even a few metres high 20–30 miles (*c.* 33–48 km) away, keep the sea continuously agitated, forming swirling currents of great velocity if stranded bergs impede the waves. They loosen and detach blocks from older bergs by rocking them, by driving them together, or by capsizing them. They generate waves which run upstream for nearly 1 mile (*c.* 1630 m) against the river, as in Alaska[34]; and they fold the sea- and fjord-ice, crack and break it up over many square kilometres, produce a schuppen structure, and hurl some of the cakes upon the ice. Wave-cut cliffs and beaches are made up to 1·5 m above the sea, the action being magnified in narrow fjords where the waves may wash lateral moraines off the hillsides or arrange flattish boulders along the shore which dip outward in imbricated fashion. With the ice and pebbles they contain, they strip off the soil, remove lichens and moss, break and kill trees,[35] and decimate intertidal communities of animals.[36] They also polish and smooth the cliffs high up above high water-mark.

The effects of such waves, which Darwin[37] stressed, may, however, be easily exaggerated, since they are local only and comparable with those of storm waves. They may be more than counterbalanced by the smoothing effect of drift bergs[38] and by fjord-ice and sea-ice which damp the waves in autumn and suppress them entirely in winter. Polar explorers who have traversed pack-ice have commented upon the tendency of sea-ice to keep the sea smooth, the undulations in the pack diminishing rapidly in amplitude from the edge[39]; the shorter period waves are the most easily damped.

Disturbances of this kind must have been important off Pleistocene west Europe and in the lateglacial sea of the Baltic. They operated in extraglacial lakes, including such as were too small to have appreciable wind-waves.[40]

Ice-foot generally protects the coast directly against waves, breakers and floating ice,[41] and indirectly by cementing the cliff talus.[42] Nevertheless, it is sometimes erosive[43] as when storm waves push its disintegrating masses up the shelving beach. It erodes by plucking[44] and by nivation along its junction with the rock[45] and aids erosion by preventing the growth of protective seaweed, barnacles or molluscs.[46]

Deposition. Erratic material carried by floe- and bay-ice and by bergs calved from ice-foot or glacier-snouts (river-ice which melts rapidly transports inappreciably[47]) is either dropped along the coast[48] or near inshore, sometimes as submerged banks,[49] or it is drifted out to sea and deposited as single erratics or as shoals, banks or islands[50] where warmer air and sea attack the ice: ice-rafting of materials has often been observed in Arctic seas[51] and in the Baltic.[52] Much of the load goes down early and within the pack-ice zone (almost pure diatom ooze occurs outside this zone in the Antarctic[53]), since coast-ice does not drift far out and ice with englacial debris melts rapidly.

Overturned bergs carry their burden farther than those which keep their original position; they may move 2500 miles (*c.* 4000 km).

Ice-foot, charged with pebbles from the beach and talus from the cliff, is sometimes an important transporting agency.[54] Yet it generally conveys little.[55] Much is above high tide and melts gradually *in situ*, especially if drift-ice protects the coasts, while that which is down the beach is frozen so firmly that it is not broken by waves but slowly disintegrates. It may give rise to terraces.[56]

The importance of the transport by drift-ice may be gauged from the fact that on an average 800 sq. miles (*c.* 2000 sq. km) of drift-ice pass any particular point on the west Greenland coast in one summer day[57] and from the vast expanse of sea over which the ice floats. This is difficult to determine but may be 21 million sq. km in the northern hemisphere and 88 sq. km in the southern hemisphere[58] (see p. 193). Greenland bergs distribute their burden over an area nine times as big as Greenland itself (blocks and gravel, for example, drift to Iceland[59]) and Antarctic bergs weather over an area four to five times larger than this continent[60]: the Antarctic is surrounded by a zone of glacial marine sediment 200–700 miles (320–1120 km) broad[61] (see below).

The Grand Bank of Newfoundland, 125,000 sq. km in extent, has been interpreted as a frontal moraine[62] or as the accumulated droppings of bergs and coast-ice.[63] The local nature of the material dredged up[64] appears to be as incompatible with the one as are the general cleanness of the bergs and lack of any berg concentration on the Bank with the other.[65] It is, indeed, doubtful whether the whole Bank was built up by drift-ice[66]; like the banks farther south, it may consist of solid rock[67] and be tectonic,[68] since it is situated at the intersection of two structural axes.

Antarctic seas are bestrewn with heterogeneous glacio-marine deposits[69] which grade into the organic ooze with no sharp contact and are characterised by poor sorting, an almost complete lack of calcium carbonate, and low organic content. The presence on the floor of Ross Sea of some rounded grains and of sericite and amorphous masses of tiny grains indicates a derivation from rocks which have contained decomposition and alteration products. The percentage of clay and especially colloid is unusually high, due to the failure of marine currents to elutriate the finer particles. Ross Sea is floored with a material which has the characteristics of a till laid down on land, the materials being distinguished by their freshness and angularity. Glacial muds, grey in colour, encircle the continent to a width of 200–500 miles (*c.* 320–800 km) and to a depth of 2000 fathoms (3660 m): they floor, for example, Bellingshausen Sea.[70]

The Arctic Ocean has in marked contrast but little drifted material,[71] save in such areas as Barents Sea and Baffin Bay; the difference is readily explained by the unglacierised northern coasts of America and Eurasia and the influx of fluviatile detritus. Fine-grained shales are being produced which give no textural indication that they ever were associated with ice or an arctic climate.[72] Bergs may even have been absent during the Glacial period since the *Fram* brought up only fine material.[73] The sediment in Davis Strait and probably off the whole coast of west Greenland is much sandier and better sorted than in the Antarctic.[74]

Marine till contains unbroken fossils and shells with valves still attached to stones upon which they lived.[75]

The geological action of drift-ice was manifestly vastly more important during the Glacial period (see ch. XL), though it did not apparently include typical boulder-clay as was at one time thought (see ch. XXX).

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